Quaternary Geology of Alaska

By TROY L. PÉWÉ

GEOLOGICAL SURVEY PROFESSIONAL PAPER 835

A study of the glacial, periglacial, eolian, fluvial, lacustrine, marine, and volcanic deposits of Quaternary age in Alaska and paleoclimatic fluctuations in light of formation and disappearance of glaciers and permafrost and changes in the distribution of plants and animals

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1975

Quaternary Geology of Alaska

By TROY L. PÉWÉ

GEOLOGICAL SURVEY PROFESSIONAL PAPER 835

A study of the glacial, periglacial, eolian, fluvial, lacustrine, marine, and volcanic deposits of Quaternary age in Alaska and paleoclimatic fluctuations in light of formation and disappearance of glaciers and permafrost and changes in the distribution of plants and animals

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1975

UNITED STATES DEPARTMENT OF THE INTERIOR

ROGERS C. B. MORTON, *Secretary*

GEOLOGICAL SURVEY

V. E. McKelvey, *Director*

First printing 1975

Second printing 1977

Library of Congress Cataloging in Publication Data

Pewe, Troy Lewis, 1918-
Quaternary geology of Alaska.
(Geological Survey Professional Paper 835)
Bibliography: p. 122-139.
Includes index.
1. Geology, Stratigraphic–Quaternary. 2. Geology–Alaska. I. Title. II. Series: United States. Geological Survey. Professional Paper 835.
QE696.P43 551.7'9'09798 75-1041

For sale by the Superintendent of Documents, U.S. Government Printing Office
Washington, D.C. 20402

Stock Number 024-001-02685-6

ILLUSTRATIONS

TABLES

QUATERNARY GEOLOGY OF ALASKA

By Troy L. Péwé

ABSTRACT

The earliest account of the geology of an Alaskan area was concerned largely with Quaternary deposits. Otto von Kotzebue in 1816 examined the famous bone-bearing frozen silt and mysterious masses of ground ice at a place now called Elephant Point in western Alaska. Alaska has long intrigued Pleistocene geologists, not only because of its interesting, and in places remarkably complete, stratigraphic record of Pleistocene events, but also because spectacular effects of the cold-climate geomorphic processes are well recorded there.

If colluvium is included, Quaternary deposits mantle virtually all of Alaska. Glacial deposits are widespread in southern Alaska and on the north and south sides of the Brooks Range. Eolian sediments are widespread in the interior and western part of the State as well as in the far north. Lacustrine sediments of middle and late Pleistocene age are exposed in the Copper River Basin and elsewhere, and fluvial deposits are extensive in major valleys and large tectonic basins.

Glaciers have covered about 50 percent of the present area of Alaska at one time or another, but large areas in central and northern Alaska have never been glaciated. The extents of late Tertiary and early Pleistocene glacial advances remain almost unknown, but those of middle and late Pleistocene and Holocene advances are well established. In Illinoian, Wisconsinan, and Holocene times, glaciers were much more extensive in southern Alaska than in northern Alaska and much more extensive on the south flanks than on the north flanks of individual mountain ranges, indicating that glaciers were nourished chiefly by air masses moving northward or northeastward from the northern Pacific Ocean.

D. J. Miller's pioneering work and the subsequent work of others in southern Alaska indicate that glaciation in Alaska started by at least late Miocene time and perhaps early middle Miocene time. Glacial deposits of Illinoian age were not commonly recognized prior to 1945, but field and photogeologic studies since then have shown that drift of Illinoian age is present in nearly every glaciated area in Alaska. Drift of Wisconsinan age is found in nearly all the mountainous areas of Alaska. Areas south of the crest of the Alaska Range and the Alaska Peninsula were almost completely inundated by ice, and icecap conditions prevailed over large areas. The Wisconsinan glacial cycle in Alaska was clearly a complex event consisting of at least two major glacial advances.

Loess probably is the most widely distributed sediment of Quaternary age in Alaska. It forms a blanket, ranging in thickness from a few millimetres to more than 60 m, that covers almost all areas lying below altitudes of 300–450 m. Thick deposits of loess are most widely distributed in central and western Alaska. Most of the loess was deposited during Illinoian and Wisconsinan time, but windblown silt is still being deposited in many areas, particularly along most braided streams in Alaska; measurements indicate an accumulation of 0.2–2.0 mm per year at present.

Extensive areas of Alaska are underlain by deposits that are the result of mass wasting or frost action in a rigorous climate—a climate widely known as periglacial. A widespread and unique phenomenon of this climate is permafrost, or perennially frozen ground, which is present throughout 82 percent of Alaska. The greatest known thickness is 650 m at Prudhoe Bay.

It is estimated that 10 percent of Alaska is covered by fluvial deposits, excluding the overlying loess blanket on the terraces. Many large tectonic basins such as the valleys of the Kuskokwim and Tanana Rivers, the Yukon Flats, and Yukon-Koyukuk lowland are filled with 1 to a few hundred metres of Quaternary fluvial sediments. Known lacustrine deposits are relatively limited in Alaska, and only one large area, the Copper River Basin, exhibits widespread, well-developed lake deposits. Alaskan marine deposits of Quaternary age now above sea level are limited to a narrow strip of land along the present coast, except in northern Alaska and perhaps in the Yukon-Kuskokwim Delta area where they are more extensive.

Volcanic ash deposits are widespread and abundant in Quaternary sediments throughout southern Alaska, from the far west in the Aleutian Islands to southeastern Alaska. These ash layers have great potential for resolving problems of Quaternary geology, archaeology, paleopedology, and palynology.

The distribution of vegetation in Alaska is divided phytogeographically into two large areas, the glaciated and unglaciated regions. The glaciated areas have a floral history of repopulation by either tundra or forest during the past 1,000 to tens of thousands of years. The unglaciated areas constitute 50 percent of the State and represent refugia for plant and animal life during glacial maximums. In Wisconsinan time, the pollen record indicates that an Arctic herbaceous tundra was present on Seward Peninsula; forest was essentially absent from the interior of Alaska until the end of Wisconsinan time. The tree line was 500–600 m lower near Fairbanks at that time.

A rich Quaternary fauna is present in Alaska, but only in the study of marine invertebrates and terrestrial vertebrates has a serious beginning been made. During the two decades since detailed studies of Quaternary stratigraphy began in Alaska, a modest but growing number of vertebrate fossils have been found in significant stratigraphic context; most are from the Fairbanks area. *Canis, Rangifer, Mammuthus, Bison, Cervus, Ochotona, Lemmus, Microtus, Equus,* and *Pliomys* are now known from sediments of pre-Illinoian age. The list of documented finds of mammal remains of Illinoian age is slowly growing, and it is evident that many taxa were present in North America earlier than is generally recognized. It is now known, for example, that *Rangifer, Ovibos, Alces, Saiga, Pliomys, Bootherium, Symbos,* and *Dicrostonyx,* as well as *Bison, Mammuthus,* and *Equus,* were in North America prior to Wisconsinan time.

Many mammals became extinct in Alaska at the end of Pleistocene time, and most of the species were grazers. It is thought that the loss of grassy habitat alone did not cause the extinction of the grazers at this time; rather, the restriction of ideal habitat plus additional stress in the form of man probably caused the demise of these species.

Climatic fluctuations during Quaternary time in Alaska were responsible for the formation and disappearance of glaciers and permafrost and changes in distribution of plants and animals. Extensive glacial advances during Illinoian time indicate a long duration of a

colder or wetter climate than now. The growth of ice wedges in central Alaska indicates a rigorous periglacial climate with a mean annual air temperature of at least −7° to −8°C, which is about 2°–4°C colder than the present. In Wisconsinan time precipitation probably did not increase markedly, but rather, mean summer temperatures decreased, and summer cloudiness increased. According to the calculated position of the past snowline, the mean July air temperature in Nome and Kotzebue during Wisconsinan time was 1.9°–2.1°C colder than today. Fairbanks and Northway were a minimum of 4°C and 4.8°C colder, with mean July Wisconsinan air temperatures of 11.5° and 9.9°C, respectively.

Frozen mammal carcasses found in Wisconsinan sediments support the concept that permafrost has existed from the time of their death until now, indicating that the mean annual air temperature has been colder than about −1°C since their death, except perhaps for short periods of time.

It is believed that in much of Alaska the climate of Wisconsinan time gradually warmed to a postglacial thermal maximum between 3,500 and 7,500 years ago. However, in central and western Alaska there is good evidence for a warm period beginning 10,000 years ago. It is as yet unclear whether evidence from the north slope represents the early postglacial warm period or the conventional thermal maximum. Glacial retreats and thermal records from permafrost indicate a warming during the last 100 years.

INTRODUCTION

It has been accurately said that the environment of the Pleistocene Epoch that formerly existed in temperate climes still exists in Alaska. Volcanic, marine, fluvial, lacustrine, glacial, and eolian deposits of Quaternary age are widespread in Alaska and are being formed today, many under periglacial conditions. Both glacial and nonglacial marine sediments are accumulating along the extensive coast, and volcanic eruptions are common in the southwestern part of the state. Glaciers are widespread. Dust is blown from active valley trains and outwash fans to be deposited as loess over adjacent terrain (fig. 1). Geologic processes, such as solifluction, frost action, and formation of permafrost, common in cold regions under periglacial conditions, are active in most parts of the State.

This paper summarizes results of studies in many areas of Quaternary geology. It is hoped that this account will lead to a better understanding of the Pleistocene and Holocene Epochs in Alaska and will also be a useful résumé of the progress of investigations in Quaternary geology in this important section of North America. A serious attempt has been made to point out current areas of disagreement, unsolved problems, and fertile areas for future research.

The following discussion is arranged according to the age of the deposits or events, insofar as information on age is available. Where the age data are too scanty, the presentation is ordered on the basis of the character and distribution of the deposits and landforms. This report was written between 1960 and 1972, and no studies published after 1972 are included, unless unpublished versions were available earlier.

GEOGRAPHY

LOCATION AND EXTENT OF AREA

Alaska is part of the largest peninsula of North America and extends from long 130°W to 173°E and from lat 52° to 72° N. (pl. 1). It is bounded on the north by the Arctic Ocean, on the south by the Pacific Ocean, on the east by Canada, and on the west by the Bering Sea, Bering Strait, and Chukchi Sea.

The area of Alaska is 1,520,000 km², about 20 percent of the area of the 48 contiguous United States. The state contains extensive lowlands and towering glacier-clad mountains, including Mount McKinley, 6,660 m in elevation, the highest peak in North America.

PHYSIOGRAPHY

The main physiographic provinces of Alaska are outlined by major topographic units and are similar to those of the western United States and Canada. The four major provinces, which were outlined by Brooks (1906) and were used in the most recent description of the physiographic divisions of Alaska (Wahrhaftig, 1965), are from south to north the Pacific Mountain System, the Intermontane Plateaus, the Rocky Mountain System, and the Arctic Coastal Plain (pl. 1). The Pacific Mountain System extends in an arc from southeastern to south-central Alaska and includes to the west the Alaska Peninsula and the Aleutian Islands. It contains the coastal mountains, Alaska Range, Wrangell Mountains, Talkeetna Mountains, Aleutian Range, and intervening lowlands. The Intermontane Plateaus include the lowlands and rolling hills of interior and western Alaska between the Alaska Range on the south and the Brooks Range on the north and also the Bering platform. The Brooks Range constitutes the Rocky Mountian System. The Arctic Coastal Plain extends north from the Rocky Mountain System to the sea. These major divisions have been subdivided into 12 provinces and further subdivided into 60 sections by Wahrhaftig (1965) (pl. 1).

CLIMATE

As might be expected, the climate of Alaska varies greatly; in fact, its range is greater than that between Florida and Maine in the conterminous United States. This is caused by a varied topography, different conditions of the seas bounding Alaska on three sides, and its great geographical extent. The northernmost point of Alaska is within 18° of the latitude of the North Pole, and the southernmost tip of southeastern Alaska is near the latitude of Copenhagen, Denmark.

Several distinct climatic zones have been recognized in Alaska, and the early climatic subdivisions (Abbé,

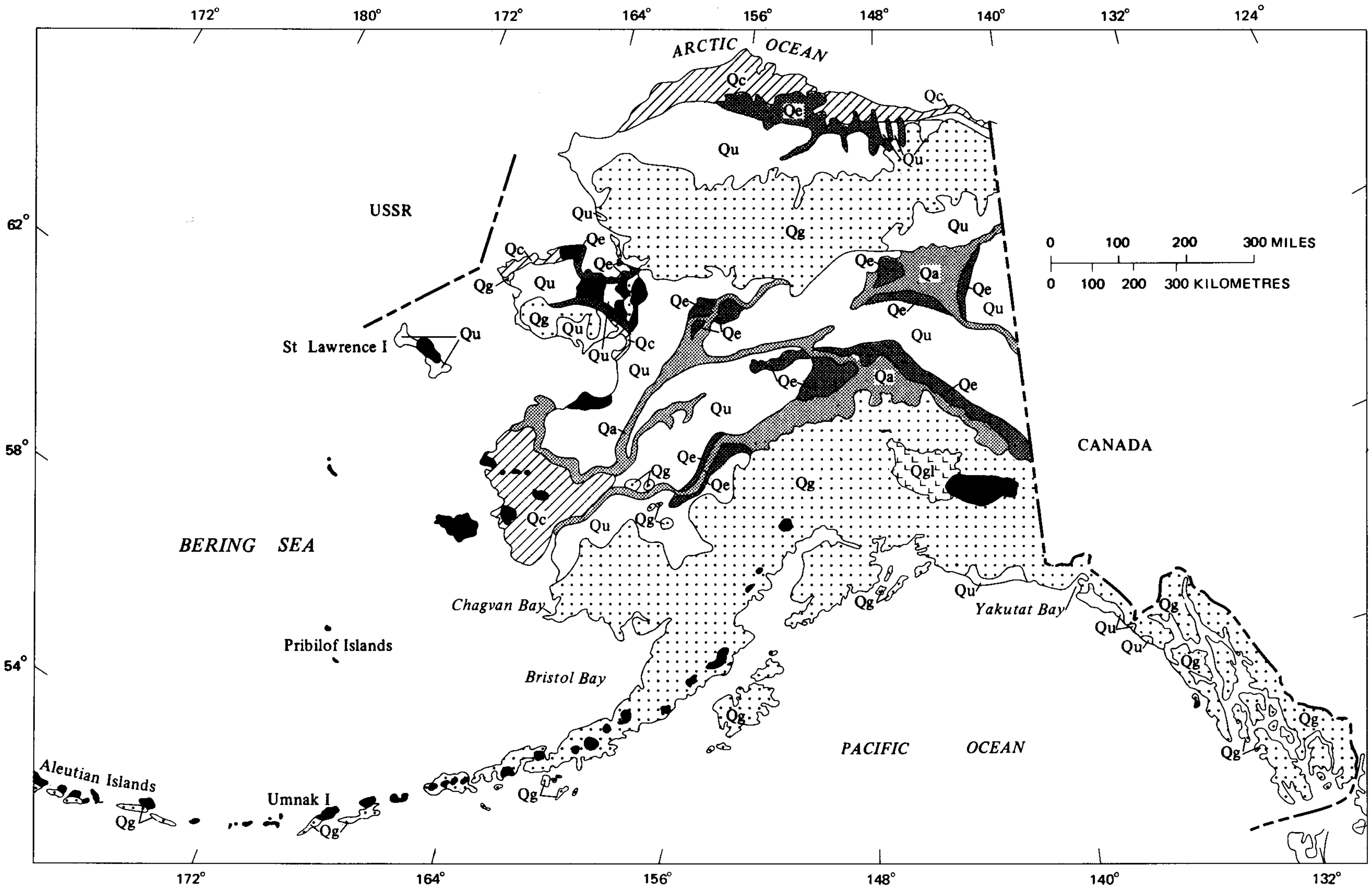


FIGURE 1.—Sketch map of major regional groups of surficial deposits in Alaska. Qg, glacial and other deposits associated with heavily glaciated alpine mountains; Qgl, glaciolacustrine deposits of larger Pleistocene proglacial lakes; Qu, undifferentiated deposits associated with generally unglaciated uplands and lowlands of the Interior and North Slope; Qa, fluvial deposits; Qe, eolian deposits; Qc, coastal deposits of interbedded marine and terrestial sediments. Solid black areas are deposits associated with volcanic peaks and flows of Tertiary and Quaternary age. Modified from Karlstrom (1960b, fig. 154.1).

1906, p. 140–141) differ only slightly from the four used today by the U.S. Weather Bureau (fig. 2) (Watson, 1959, p. 24). The southernmost climatic zone, a zone of dominant maritime influence, includes southeastern Alaska, the south-central coast, the southeastern coast of the Alaska Peninsula, Kodiak Island, and the Aleutian Islands (fig. 2). This area is characterized by small temperature variations, high cloudiness, and abundant precipitation, especially in southeastern Alaska. The highest recorded mean annual precipitation, 560 cm, is at Little Port Walter in southeastern Alaska. This zone has no permafrost but contains extensive glaciers.

A zone of transition from maritime to continental climate lies north of the maritime zone (fig. 2). This zone includes the Wrangell Mountains and extends to the west, widening to include the Cook Inlet area, the southern Alaska Range, the Yukon-Kuskokwim Delta, and the southern half of the Seward Peninsula. Temperature variations in this zone are more pronounced, and there is less cloudiness and precipitation than in the maritime zone. The mean annual air temperature is colder that 0°C (fig. 3), and permafrost is common in this zone.

The dominant continental climate zone lies north of the Alaska Range, south of the Brooks Range, and east of the Seward Peninsula (pl. 1). These topographic barriers and remoteness from the closest open ocean areas tend to prevent the inland movement of moist maritime air masses; the area is semiarid. This zone has great extremes of temperature, from 37.8°C to −60°C. The mean annual air temperature ranges from −4°C to −8°C (fig. 3), and permafrost is widespread.

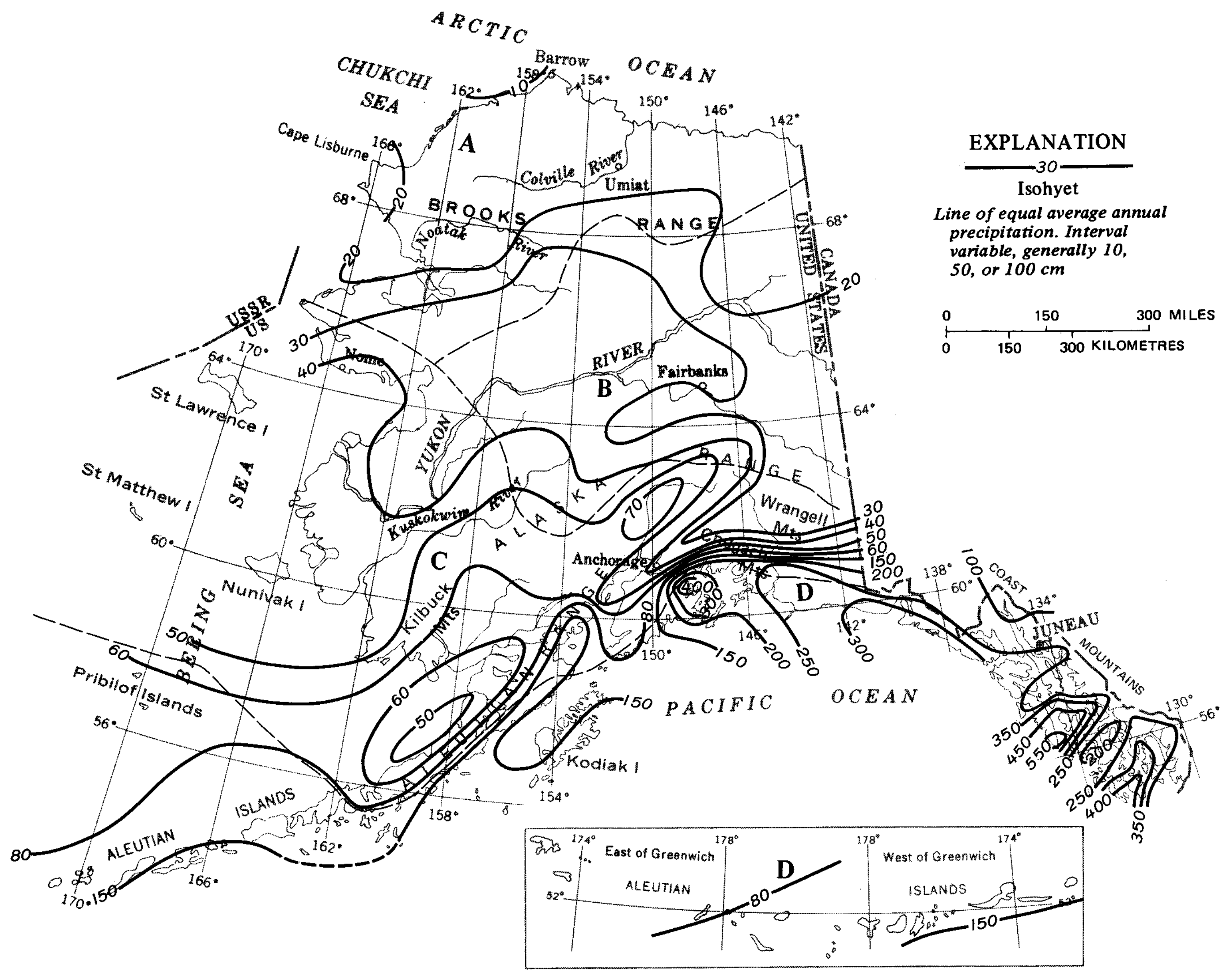


FIGURE 2.—Mean annual precipitation (centimetres) in climatic zones. A, Arctic; B, Continental; C, Maritime-Continental Transition; D, Maritime. Modified from Watson (1959, p. 19).

In the zone of Arctic climate, in northern and northwestern Alaska including the north half of the Seward Peninsula (fig. 2), mean annual air temperature ranges from −4°C to −12°C, and precipitation is slight (10–30 cm annually). Permafrost is present almost everywhere.

ACKNOWLEDGMENTS

The writer wishes to express deep appreciation to D. M. Hopkins, who assisted with the organization of the report and reviewed the manuscript. Others who reviewed and made valuable suggestions and contributions to the sections dealing with their specialties are Jerry Brown, R. L. Detterman, O. J. Ferrians, Jr., A. H. Lachenbruch, D. S. McCulloch, M. F. Meier, D. R. Nichols, George Plafker, A. S. Post, C. E. Ray, C. A. Repenning, P. V. Sellmann, Dwight Taylor, F. R. Weber, and J. R. Williams. Their help is greatly appreciated. L. R. Mayo and R. D. Reger assisted in the compilation of many of the illustrations and tables and in the preparation of the manuscript.

HISTORY OF RESEARCH

The earliest account (Kotzebue, 1821) of the geology of an Alaskan area was concerned largely with Quaternary deposits. Kotzebue's report contains an account of the famous bone-bearing frozen silt and of mysterious masses of buried ice at a place now called Elephant Point on Eschscholtz Bay, about 35 km south of the Arctic Circle. Accounts of exploration in Alaska during the latter part of the 19th century and the many accounts of the regional geology published by the Geologi-

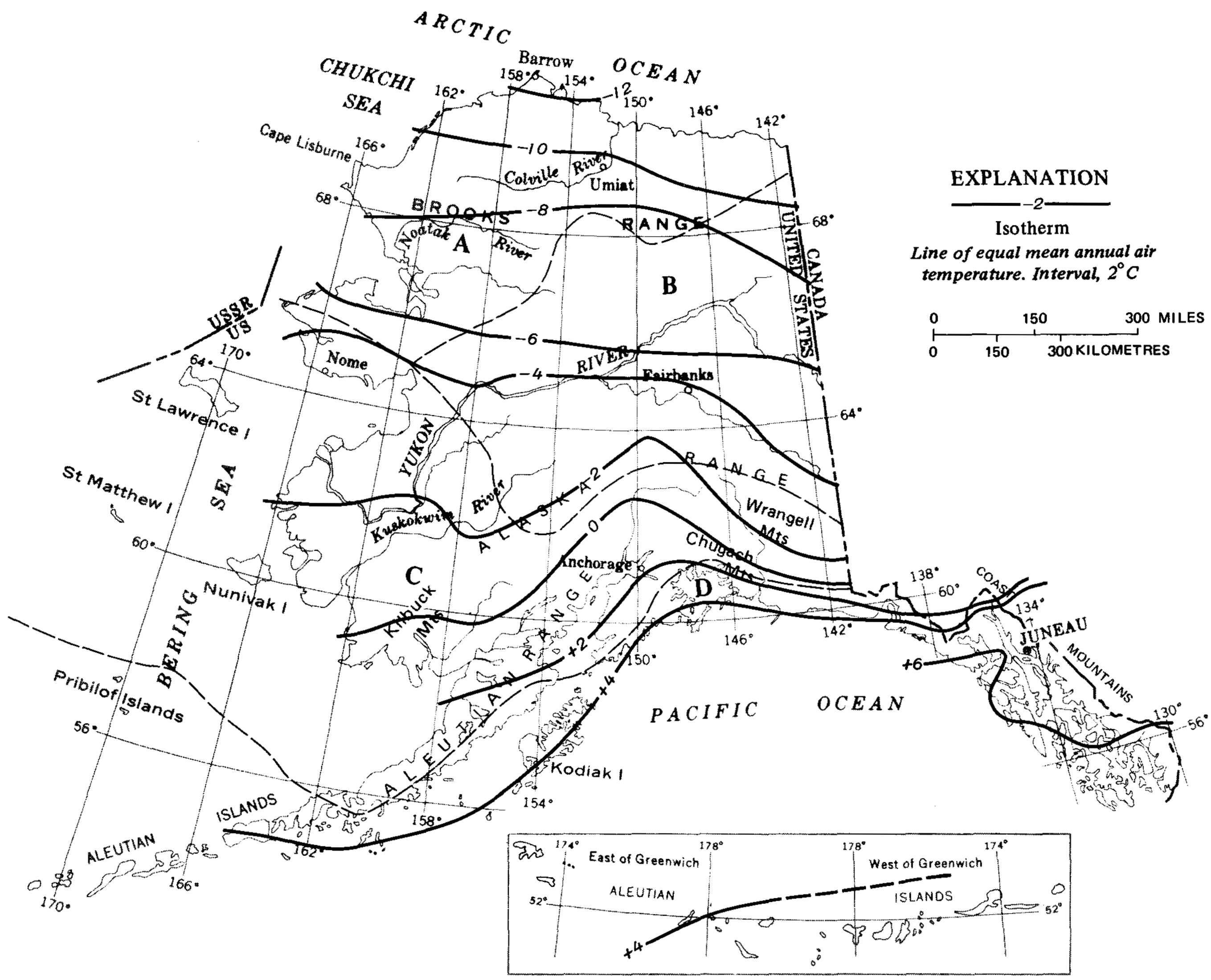


FIGURE 3.—Mean annual air temperature (degrees Centigrade) from Péwé (1966b, fig. 4; effect of topography not considered). Climatic zones: A, Arctic; B, Continental; C, Maritime-Continental Transitional; D, Maritime. Modified from Watson (1959, p. 24).

cal Survey from 1898 through 1940 almost invariably contain descriptions of the physiography and of the Quaternary sediments as well as descriptions of the bedrock geology. Brooks (1906) wrote the earliest comprehensive account of Alaska Quaternary geology.

GLACIOLOGY

The magnificent and spectacularly fluctuating glaciers of southern Alaska (fig. 4) began to receive attention in the latter half of the 19th century, through the work of Muir (1884), Reid (1892, 1896), and Russell

FIGURE 4.—Oblique aerial view of Columbia Glacier entering Prince William Sound, south-central Alaska. View to the north with Chugach Range in the background. The highest peak is Mount Witherspoon, 3,644 m in elevation. Columbia Glacier has an area of 960 km^2. Photograph by Austin S. Post, U.S. Geological Survey, August 25. 1965.

(1892a, 1892b, 1893). Coastal Alaska's great glaciers have always been of interest to Pleistocene geologists and have been under close observation since the 1890's (Russell, 1892a; Reid, 1896; Gilbert, 1910). Tarr and Martin (1914) summarized their systematic observations, and Grant and Higgins (1913) carefully observed the glacier termini in Prince William Sound and Kenai Peninsula. Continuing studies between 1920 and 1950 (Cooper, 1937; 1939; Field, 1932a, 1932b, 1941, 1942, 1947, 1950; Wentworth and Ray, 1936) led ultimately to coordinated glaciological and glacial geological studies of the Malaspina-Seward ice field system (Sharp, 1951a, 1951b, 1956, 1957) and of the Juneau ice field (Miller, 1952, 1953, 1970; Field and Miller, 1950, 1951; Hubley, 1955, 1957; Thiel and others, 1957; Heusser and Marcus, 1964). The glaciers of interior Alaska received less intense scrutiny for many years, although the spectacular advance of Black Rapids Glacier in 1937 (Hance, 1937; Moffit, 1942; Geist and Péwé, 1957) and Muldrow Glacier (Péwé, 1957a; Post, 1960) indicated that interesting problems awaited study. Only recently, with the growth of the study of glacier surges (Meier, 1969; Post, 1969; Meier and Post, 1969), has it been understood that Black Rapids Glacier and Muldrow Glacier advances were the vanguard of documented glacial surges.

Systematic glaciological and glacial geological studies in the Alaska Range and in the Brooks Range did not begin until the International Geophysical Year in 1957, with the studies of Péwé and his associates in the Alaska Range, mainly on Gulkana Glacier, (Péwé, 1957a, 1957b, 1961a, 1961b, 1963; Sellmann, 1962; Moores, 1962; Mayo and Péwé, 1963; Mayo, 1963; Reger, 1964, 1968; Ragan, 1964, 1966, 1967, 1969; Rutter, 1965; Ostenso and others, 1965; Reger and Péwé, unpub. data, 1963) and of R. C. Hubley, G. W. Holmes, and E. G. Sable in the Brooks Range (Keeler, 1958; Sater, 1959; Holmes and Lewis, 1961; Sable, 1961). Also, pioneer glaciological studies in the Wrangell Mountains were published (Benson, 1968). The U.S. Geological Survey now has a long-range detailed program of glacier studies on Gulkana Glacier in the Alaska Range and on Wolverine Glacier on the Kenai Peninsula (Meier and others, 1971).

The Alaska earthquake of March 27, 1964, provided an unequaled opportunity to examine the effect of an earthquake on glaciers, because the epicenter was in Prince William Sound, an area bordered by hundreds of glaciers. Many observations and studies were made of glaciers and of avalanches on glaciers during the first days and months after the quake (Bull and Marangunic, 1968; Field, 1965, 1968; Field and others, 1964; Hackman, 1965; La Chapelle, 1968; Marangunic and Bull, 1968; Marcus, 1968; Miller, 1965a, b; Péwé, 1964a, b, 1968b; Plafker, 1968; Post, 1967; Ragle and others, 1965a, b; Reid, 1968; Sater, 1964; Shreve, 1966, 1968; Tuthill, 1966; Tuthill and others, 1968).

As a result of the 1899 earthquake in southern Alaska and a study of glacier advances there, Tarr and Martin (1914) discussed the ideas that earthquakes caused glaciers to advance either immediately or at a later date as a result of great quantities of snow on mountainsides being shaken loose and cascading down on the glaciers below. After the 1964 quake observations from the air of hundreds of glaciers revealed no advancing ice streams as the result of the earthquake. It will take some time, however, to determine whether the glaciers will expand as the result of snow avalanched down onto the ice. Observations indicate that most of the snow knocked down would have fallen later in the spring under ordinary circumstances. Spectacular debris avalanches (fig. 5) on glaciers did occur, and studies are underway to evaluate the effect of the debris on glacier regimen (Bull, 1969).

GLACIAL STRATIGRAPHY

Aside from attempts to date Holocene fluctuations of the coastal glaciers, studies of glacial stratigraphy were for many years quite desultory. To be sure, most Geological Survey reports took note of the distribution of moraines and glaciated terrain; Capps particularly concerned himself with the possibility that more than one glacial cycle was represented in Alaska and summarized (1931) the existing knowledge of the glacial history of Alaska. Most of the recognized glaciated areas rather clearly dated from the glaciation of Wisconsinan age (Capps, 1915b), and older glacial deposits were commonly overlooked; however, Capps (1916b) discovered deformed deposits of apparent glacial origin and of probably early Pleistocene age in the White River area on the north and east flanks of the Wrangell Mountains (Capps, 1910b, 1915a), and Taliaferro (1932) discovered marine tillites of late Tertiary and possible Pleistocene age along the Gulf of Alaska; these were studied in more detail by Miller (1953a, 1957). Shortly after World War II, intensive studies of glacial stratigraphy were begun in many parts of Alaska and resulted in a brief publication by Péwé and others (1953). Several notable papers on individual areas include those by Wahrhaftig (1958), Detterman and others (1958), Miller and Dobrovolny (1959), Fernald (1960), Hopkins and others (1960), Holmes and Lewis (1961), Coulter and Coulter (1961, 1962), Péwé and Holmes (1964), and Karlstrom (1964). These and other works were summarized by the Alaska Glacial Map Committee of the U.S. Geological Survey in a map showing the extent of glaciations in Alaska (Coulter and others, 1965).

Péwé, Hopkins, and Giddings (1965) summarized various events, and Péwé, Ferrians, Nichols, and Karlstrom (1965) presented detailed information on Quaternary geology in central Alaska. Detailed work on glacial deposits of the Brooks Range was only recently undertaken (Fernald, 1964; Porter, 1964, 1966; Holmes and Lewis, 1965; Hamilton, 1969; Reed, 1968). The following glacial mapping elsewhere in Alaska adds a fuller understanding of the glacial history of the State: The upper Tanana valley (Fernald, 1965a), York Mountains (Sainsbury, 1967a), Chagvan Bay area in southwest Alaska (Porter, 1967), the Johnson River area of the Alaska Range (Holmes and Foster, 1968), and the Ray Mountains (Yeend, 1971). The chronology of the late Holocene in Alaska was well summarized by Porter and Denton (1967). Rampton (1970) reported recent glacial changes near the Alaska-Canada line.

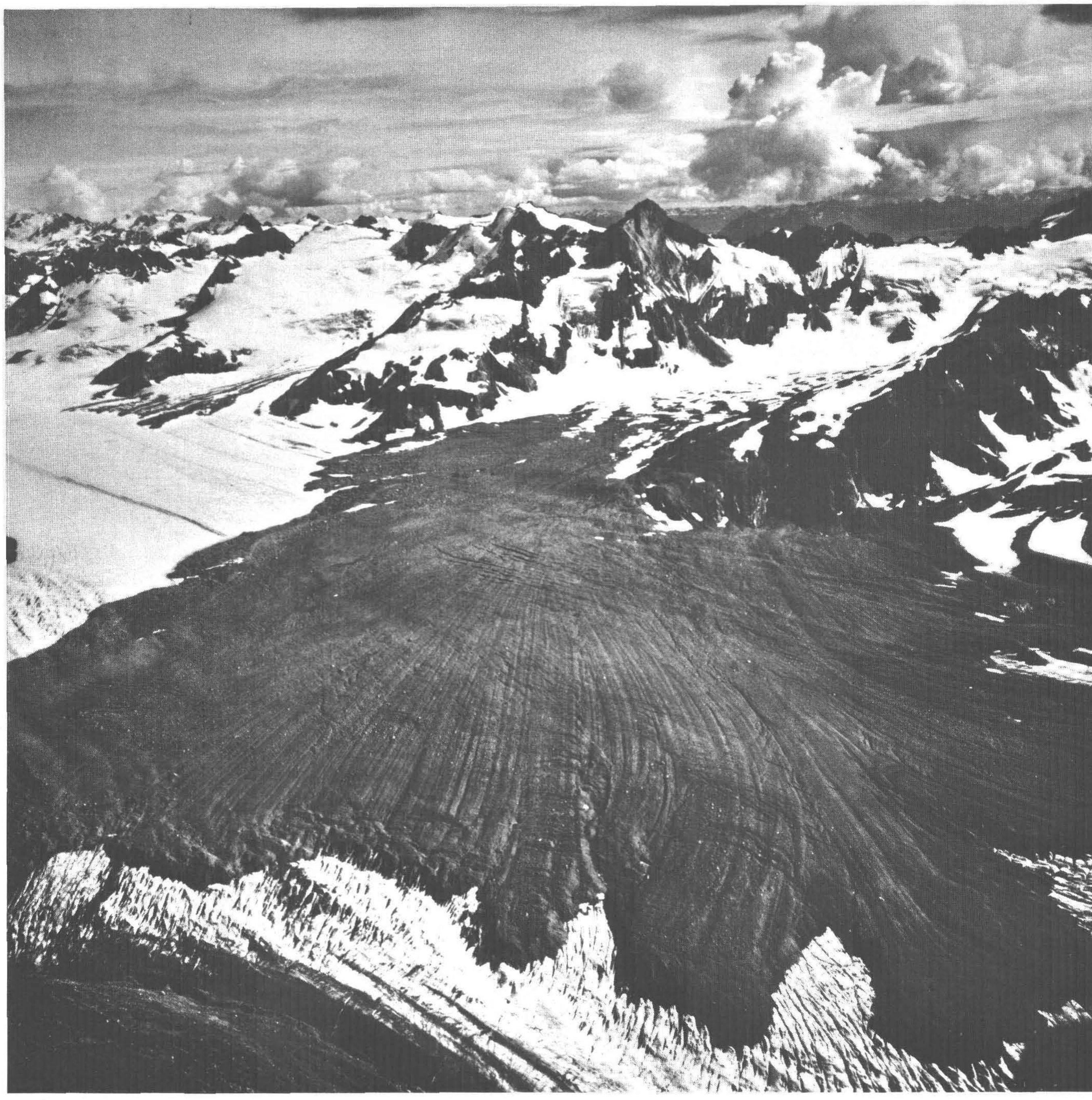


FIGURE 5.—Earthquake-induced landslide of 1964 on Sherman Glacier, south-central Alaska; $10^7 m^3$ of slide debris covers 8 km^2 of ablation zone. Glacier regimen changed from negative to strongly positive (Marangunic and Bull, 1968). Photograph by Austin S. Post, U.S. Geological Survey, 1965.

Glacial mapping is being continued on the south flank of the Brooks Range by T. D. Hamilton and his associates at the University of Alaska. Denton and others are unravelling the interbedded volcanic and glacial deposits in the Wrangell Mountains with the aid of potassium-argon dating (Denton and Armstrong, 1969; Stuiver and others, 1969). The great challenge of glacial geology today in Alaska is to relate the glacial record to the volcanic deposits by radiometric dating and paleomagnetic observations. Alaska abounds with areas in which these methods can be employed.

NONGLACIAL DEPOSITS

The stratigraphy and paleontology of the richly fossiliferous silts of nonglaciated areas of Alaska continue to arouse interest that began with the earliest studies by the Kotzebue and Beechey expedition. In 1907 and 1908, Quackenbush (1909) revisited and reexcavated the historic bluff at Elephant Point. Eakin (1916; 1918) and Mertie (1937) gave extended accounts of silt deposits in central Alaska. Taber (1943) presented an even more elaborate account of the stratigraphy of silts near Fairbanks that he considered to be the result of long-continued frost-riving of bedrock; however, Tuck (1940) had shown that these deposits were actually loess, and Péwé (1955) confirmed this conclusion and also demonstrated a long and complex late Pleistocene history in the Fairbanks area. The University of Alaska in cooperation with the American Museum of Natural History has collected vertebrate remains from the placer pits near Fairbanks, and in other parts of Alaska through the efforts of Geist (1953), an enormous quantity of vertebrate material has been accumulated (Frick, 1930, 1933, 1937) of which only the bison have been systematically described (Skinner and Kaisen, 1947). Guthrie and his associates at the University of Alaska initiated a new era of study of Pleistocene vertebrates in Alaska, especially a study of the great collection from the Fairbanks area (Guthrie, 1966a, b, c; 1967; 1968a, b; 1972). The Quaternary geology of unglaciated central Alaska has been under study by Péwé (1952a; 1958a, b; 1965a, b; 1970a, b; 1975).

Understanding of the climatic history recorded in the nonglacial deposits has been deepened greatly by palynological studies conducted during the last two decades by many investigators, most notably by Heusser (1960, 1966), Livingstone (1955, 1957), Colinvaux (1962, 1963, 1964a, b; 1967a, b, c), and Matthews (1968a, 1970).

ARCHAEOLOGY

The latest events of Quaternary time involve the coming of man. In addition to enlarging our knowledge concerning the record of man, archaeological studies in Alaska have contributed greatly to our understanding of sedimentation, climatic and vegetation changes, and stratigraphy of nonglacial deposits of the last 8,000–10,000 years. A summary of the archaeology of Alaska is beyond the scope of this report, but a brief review of such research is presented because it concerns the Quaternary record.

Alaska is almost surely the route by which the first humans came to North America, but until the recent discoveries of Giddings (1962), no undisputed occupation sites or artifacts older than 6,000 years had been found there. Until the early 1940's, in fact, the only known archaeological sites aside from the Campus site (Rainey, 1939) were those occupied within the last few centuries by Eskimos. The discovery of the Ipiutak site at Point Hope, occupied about 2,000 years ago, marked a tremendous advance in our understanding of Eskimo prehistory (Larsen and Rainey, 1948). The reports of Geist and Rainey (1936) and Collins (1937) on St. Lawrence Island presented a nearly complete understanding of the last 2,000 or 3,000 years of Eskimo history there. More recently, the discovery of the Denbigh Flint complex at Iyatayet, on Norton Sound (Giddings, 1951, 1964, 1967; Hopkins and Giddings, 1953), and of 6,000-year-old artifacts in the Trail Creek caves on Seward Peninsula (Larsen, 1951, 1953, 1968) indicated that the Eskimo has had a long history in Alaska. Research on Anangula Island in the Aleutian Islands produced insight into about 8,000 years or more of Aleut-Eskimo history there (Laughlin, 1963). Radiocarbon dates of 8,000 and 8,500 years are reported, but Black and Laughlin (1964) believed the age may be more easily 12,000 years. In more recent years, many workers have found sites comparable in antiquity to the Denbigh Flint complex in other parts of northern and western Alaska (Campbell, 1962; Hadleigh-West, 1963). The present knowledge of Eskimo archaeology is ably summarized by Giddings (1960) and more recently by Dumond (1965). Shortly before his death, Giddings compiled a succinct summary of archaeology in Alaska in (Péwé, Hopkins, and Giddings, 1965).

The record of man in Alaska is constantly being pushed back in time, and latest views involve one of the oldest known sites and one of the newest. Discovered in 1933 on the campus of the University of Alaska (Rainey, 1939, 1941), the Campus site, with its typologically old artifacts with Old World relations is of major importance. Clear stratigraphy and dating are missing. Recently, Hadleigh-West (1967) described artifacts from the Donnelly Ridge site on a moraine of Wisconsinan age (Péwé and Holmes, 1964) 120 km southwest of the University of Alaska. He believed that the artifacts in the Donnelly Ridge site have many affinities with the Campus site artifacts. Because of affinities of the artifacts of both sites with Asian finds 10,000–15,000 years old, Hadleigh-West believed the Alaskan artifacts fall into this range of time.

There is no geologic reason to believe the Campus site cannot be this old, or older. The Donnelly Ridge site may perhaps be this old but not more, because the Donnelly moraine is not more than 15,000 years old. Recent discoveries by McKennan and Cook (1968) at Healy Lake 150 km southwest of Fairbanks show cultural layers older than 11,000 years, and materials similar to the Campus site seem to occur in a level dated at 8,960 years old (Hosley, 1969).

After more than 20 years of patient searching, which revealed outstanding finds, J. L. Giddings discovered in 1961 "what is undoubtedly the most important archeological site ever found in the Arctic" (Collins, 1967). This was the Onion Portage site on the Kobuk River (pl. 1; Giddings, 1962, 1967; Anderson, 1968, 1970), a clearly stratified site of silt and sand of eolian and fluvial origin that is as much as 6 m thick and contains many occupational levels. The oldest dated level is 8,500 years old, and the record may go back 6,500 years more, making it the oldest site in Alaska as well as the one with the longest continual record.

With a growing body of evidence that man was present in North America at least as early as 25,000–30,000 years ago or more, it is imperative to search further for man in Alaska in deposits of late Quaternary age.

MARINE DEPOSITS

Pleistocene and Holocene marine deposits are found in many areas on the coasts of Alaska and were recorded (though erroneously assigned to the Miocene Epoch) early in the history of exploration of Alaska by Dall and Harris (1892). The most comprehensive accounts of the Pleistocene marine mollusks are those of MacNeil and others (1943) and MacNeil (1957); Black (1964) and O'Sullivan (1961) studied the stratigraphy of the Pleistocene marine Gubik Formation of the Arctic Slope; Hopkins devoted much attention to the Pleistocene sequence at Nome (Hopkins and others, 1960); and Miller (1953b) presented the most complete account of the late Cenozoic marine deposits along the Gulf of Alaska coast. Twenhofel (1952) described the deposits of the postglacial marine transgressions in southeastern Alaska. Present knowledge of the stratigraphy and age of Quaternary marine deposits in Alaska is summarized by Hopkins (1965, 1967b), and correlation with marine deposits in the U.S.S.R. across Bering Strait has been presented (Hopkins and others, 1965). McCulloch (1967; see also McCulloch and others, 1965) studied marine Pleistocene deposits along the northwest coast as far south as Seward Peninsula, with special emphasis on the Kotzebue Sound area.

PERIGLACIAL STUDIES

Alaska has intrigued Pleistocene geologists not only because of its interesting, and in places remarkably complete, stratigraphic record of Pleistocene events, but also because the spectacular effects of cold climate geomorphic processes are well recorded there. Many of the early Geological Survey Bulletins on Alaska contain discussions of the surface features resulting from solifluction and intense frost action. Eakin (1916) and Cairnes (1912a, b) called attention to the unique aspects of the landscape cycle in regions where frost action is intense. Porsild (1938) described ice laccoliths in tundra regions of Alaska and christened them pingos. More recently, pingos were discovered and dated in central Alaska (Krinsley, 1965; Holmes and others, 1966, 1968). Sharp (1941, 1942) described microrelief features in the Wolf Creek Range just east of the Alaskan border in Canada. During the late 1940's and early 1950's interest quickened in frost features and the resulting landscape. Hopkins and Sigafoos (Hopkins and Sigafoos, 1951, 1954; Sigafoos and Hopkins, 1952) wrote a series of papers dealing with the interrelations of frost action and vegetation. Wallace (1948), Hopkins (1949), Anderson and Hussey (1963), and Black (1969b) described the origin and development of thaw lakes. Black and Barksdale (1949), Livingstone (1954), Carson and Hussey (1959, 1960a, b, 1962, 1963), and Black (1969b) offered explanations for the pronounced orientation shown by thaw lakes in many parts of Alaska. Capps (1910a) coined the word "rock glacier" for glacierlike tongues of rock debris that are common in mountainous areas in Alaska; Wahrhaftig and Cox (1959) thoroughly described rock glaciers, the mechanisms by which they move, and the climatic conditions that control their distribution.

Most writers discussing frost action and the development of periglacial landscapes have been concerned with tundra landscapes, but Drury (1956) wrote an interesting and widely quoted account of the geomorphic and vegetational processes involved in the development of "bog flats," a conspicuous variety of muskeg that is widespread on gently sloping surfaces in the forested region of central Alaska.

Leffingwell gave the first description of ice wedge polygons in Alaska and the first correct explanation of their origin in 1915. Taber (1943) contributed greatly to our understanding of frost heaving and the origin of small lenses of ground ice, but his explanation for the origin of ice wedges is no longer accepted by most workers. The investigations of Black (1951b, 1952a, 1969a), Péwé (1952a, 1958a, 1962, 1966b, 1973, 1974a, b, 1975), Lachenbruch (1960a, b, 1961, 1962, 1966), and Brown (1967a) considered the structure, petrography, stratigraphy, and thermal environment of ice wedges. Their distribution and climatic significance were noted by Péwé (1964c, 1966a).

Dissemination of information on the periglacial features in Alaska was stimulated by the field conference and symposium held in central Alaska in 1965 (Péwé, Ferrians, Nichols, and Karlstrom, 1965).

Among the least known, but most widespread, periglacial features in the world are altiplanation terraces. Study of the origin, age, and distribution of these features in central Alaska has begun (Péwé and Reger, unpub. data 1969; Péwé, 1969b, 1970b; Reger and Péwé, unpub. data 1974).

VOLCANIC ACTIVITY

The presence of active volcanoes in the Aleutian Islands and on the Alaska Peninsula was known to the earliest explorers of Alaska. An eruption on Bogoslof, a minute volcanic island north of the Aleutian chain, was described by Dall (1884, 1885). The cataclysmic, caldera-forming eruption of Mount Katmai and the resulting deposits were described in detail by Griggs (1922, and several later publications) and Fenner (1920, 1923, and many later publications). The discovery of Aniakchak caldera by Smith (1925) indicated that caldera-forming eruptions took place elsewhere in the region. An eruption at a small cone in Okmok caldera near the military post at Fort Glenn, Umnak Island, in 1945 and an eruption of Mount Cleveland on Chuginadak Island in 1944 demonstrated potential threat of the volcanoes in the Aleutian chain to human activities and resulted in a coordinated study by the Geological Survey of the geology of the entire region during the late 1940's and early 1950's (U.S. Geological Survey, 1955–61).

Basaltic lava plateaus that occupy large areas in central and western Alaska constitute a separate and different sort of volcanic province from the andesitic volcanoes of southern and southwestern Alaska. Barth (1956) and Hopkins (in Cox and others, 1966) prepared detailed descriptions of basaltic volcanic rocks on the Pribilof Islands, and Hopkins (1963) on central Seward Peninsula. Geologic mapping, paleomagnetic stratigraphy, and potassium-argon dating were used in a thorough study to determine the time and volume relations of tholeiitic and alkalic basalt on Nunivak Island (pl.1) (Hoare and others, 1968). A summary of the distribution and importance of volcanic ash deposits is included in the present report, but no summary is attempted of the Quaternary volcanoes and flows.

PALEOMAGNETIC STUDIES

Quaternary paleomagnetic stratigraphy in Alaska is in its infancy, but significant contributions have already been made. The first confirmation of the Olduvai normal polarity event was made in a study of basalt flows on St. Paul Island (Cox and others, 1966), and the Nunivak normal polarity event was established on the basis of studies on Nunivak Island (Cox and Dalrymple, 1967; Hoare and others, 1968). With further examination of the abundant widespread volcanic rocks of Quaternary age in Alaska and sediments in the Arctic Basin, great opportunities will be present to tie into, and perhaps refine, the geochronometric time scale of geomagnetic polarity epochs (Cox and others, 1963a, b, 1964a, b, 1965; Cox, 1969; Clark, 1969).

QUATERNARY TECTONISM

The first summary of Quaternary tectonics in any large part of North America was by King (1965). No such summary is available for Alaska, or perhaps yet possible in any detail. More than half of Alaska is seismically active today and has a rich tectonic record. Effects of Quaternary tectonism have long been known and are reflected in faulting of glacial deposits (Richter and Matson, 1971; Stout and Brady, 1972; Stout and others, 1972), uplifted glacial terraces in mountains (Wahrhaftig, 1958), uplifted marine terraces in south-central Alaska (Twenhofel, 1952; Miller, 1953a), the York terrace on western Seward Peninsula (Sainsbury, 1967a, b), and faults in the frozen nonglacial deposits in central Alaska (Péwé, 1958b; Patton and Hoare, 1968) and elsewhere.

The study of Quaternary tectonics received a tremendous impetus from the great Alaska earthquake of 1964 (Plafker, 1965; Wood, 1966; National Academy of Sciences, 1968, 1970, 1971). Various aspects of Quaternary tectonism are discussed by Plafker and Rubin (1967).

THE PLIOCENE-PLEISTOCENE BOUNDARY

The position of the worldwide Pliocene-Pleistocene boundary is currently a controversial issue. The problem was reviewed by Flint (1971, p. 381–382) and can be summarized as follows. Historically, the boundary was placed on the basis of either evolutionary differences between fossil organisms or evidence for the climatic cooling (Flint, 1965), expressed mainly in glaciation. In fact, the Pleistocene became equated with the glacial or the ice age period. It has been known from Alaska since 1932 (Taliaferro, 1932) that glacial deposits in the Gulf of Alaska were Pliocene in age, and this was carefully documented by Miller (1953a, 1957). However, not until the sixties did widespread evidence appear that glacial sediments, both terrestrial and marine, were deposited in Pliocene and Miocene time (table 1), especially in the high latitudes.

The presence of cold-water marine and terrestrial faunas in lower latitudes, representing a migration from the polar areas, is thought by many to mark the

TABLE 1.—*Position of the Pliocene-Pleistocene boundary*

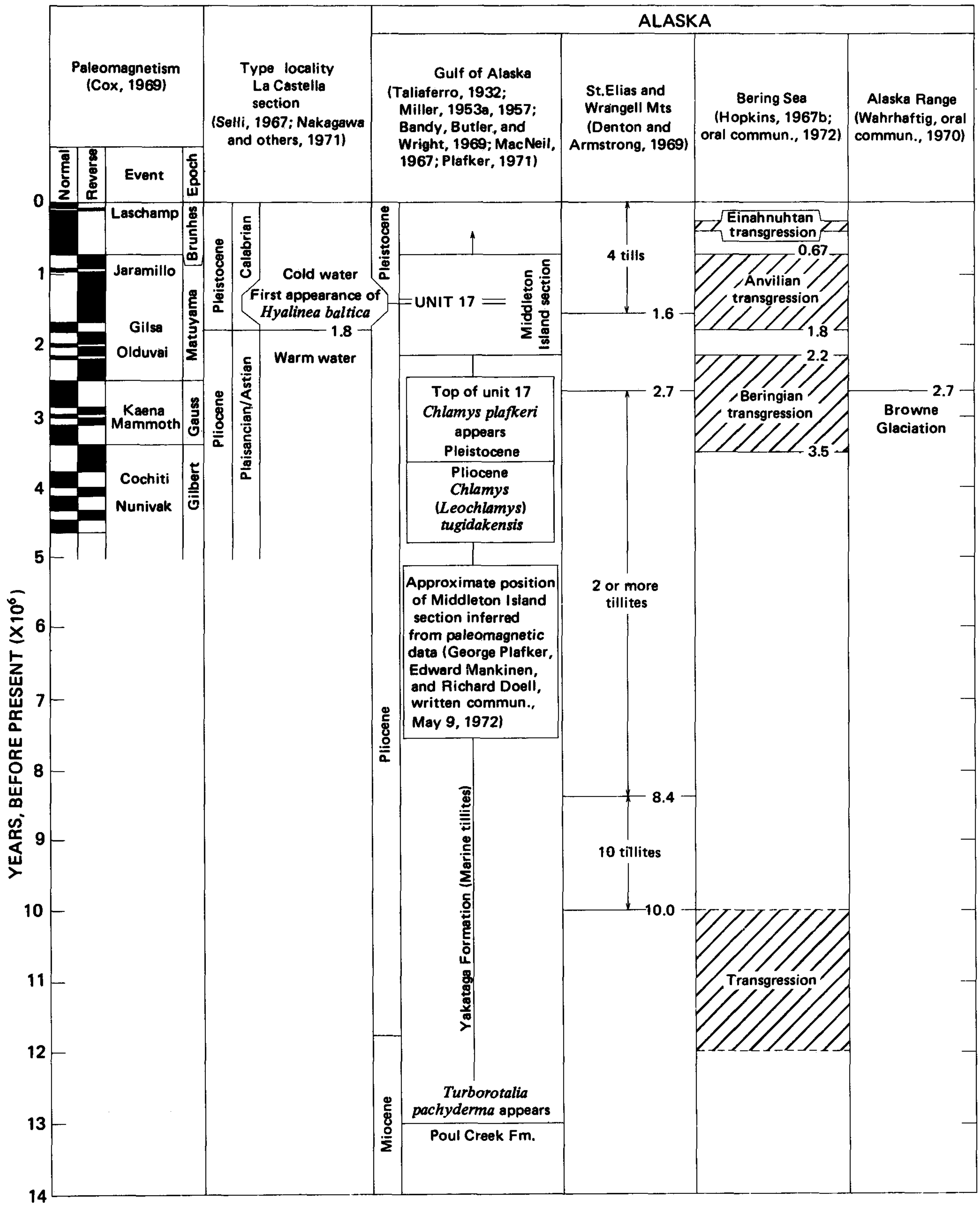


TABLE 1.—*Position of the Pliocene-Pleistocene boundary*—Continued

ALASKA						Antarctica (Denton and others, 1971)	Iceland Tjörnes Peninsula (Einarsson and others, 1967)	
Seward Peninsula		Arctic Ocean			Nunivak Island paleomagnetism (Hoare and others, 1968)			
50 km south of Cape Deceit (Hopkins and others, 1971)	Nome (Hopkins, 1973)	(Stevenwald and others, 1968)	(Hopkins and others, 1971)	(Herman, 1970; Herman and others, 1971; Herman, oral commun., Dec. 1973)				
5.7 Basalt flow Hoogendorn flora (forest of spruce, birch, pine, and hemlock)	Glaciation (inferred date) 1.5	Ocean no warmer than present 1.5 Ice rafting activity began Glacial-marine 4.5 Brown lutite	Permanent ice pack unchanged for last 10^5 years Possibly warmer water and ice rafting	Intermittent permanent ice pack 0.7 Pack ice free 2.4 (ice rafting)	Qkn1 Qmn1 Qkn2 Qmn2 QRr1 Qmr1 Qmn3 Tbn1 Tb Ta	Taylor I 0.3 Taylor II 1.6 Taylor III 2.1 Taylor IV 3.5 Taylor V Large ice sheet Ice sheet in West Antarctica	Interbedded tillite and normal lava flows of Búrfell and Grosafjöll Breidavik beds—tillites I Furuvik beds—tillites II Tjörnes beds: *Cardium* zone (first appearance of Pacific boreal mollusks)	0 1 2 3 4 5 6 7 8 9 10 11 12 13 14

Compiled by T. L. Péwé

beginning of Pleistocene time; however, such migrations are also represented in strata of late Tertiary age. Therefore, glacial climates are not a satisfactory method for determining the position of the Pliocene-Pleistocene boundary. As Flint remarked (1971, p.2), we now think in terms of late Cenozoic cold climates and glaciers and do not associate them only with the Quaternary. In a rather informal manner, some scientists use the term Quaternary to mean the last 2 or 3 m.y. (million years).

Under the auspices of international congresses, the position of the Pliocene-Pleistocene boundary (and therefore the Tertiary-Quaternary boundary) has received much study and official recommendations have been made. It was recommended at the 18th International Geological Congress in 1948, the 6th International INQUA Congress in 1961, and at the 8th International INQUA Congress in 1969 that the La Castella section in southern Italy be regarded as the boundary stratotype fixing the position of the Pliocene-Pleistocene marine boundary. The boundary is placed at the horizon characterized by the sudden appearance of *Hyalinea baltica* (table 1), a cold-water species of Foraminifera which migrated from the North Atlantic. Selli (1967) and Nakagawa, Niitsuma, and Elmi (1971) gave a date of 1.8 m.y. for the Pliocene-Pleistocene boundary in the La Castella section (table 1) on the basis of magnetic, microfaunal, and radiometric evidence.

In this report, the writer uses the date 1.8 m.y. for the temporal position of the Pliocene-Pleistocene boundary, although it is abundantly evident that in Alaska there is no known sharp break at this time in the glacial record, in either marine or terrestrial sediments, or an abrupt change in faunal or floral types (table 1).

Attention has only recently been focused on the problem of defining the Tertiary-Quaternary boundary in Alaska. Continuing studies of faunas, floras, and remnant magnetism coupled with potassium-argon dating of rocks of late Tertiary or early Quaternary age have begun to provide more precise age assignments (table 1). In many areas, sediments of middle or late Pleistocene age lie unconformably on much older, pre-Quaternary rocks; in other areas the oldest Quaternary rocks whose ages can be established rest on still older, little deformed sediments or volcanic rocks of either early Quaternary or late Tertiary age.

At the present time, the position of the Pliocene and Pleistocene boundary can be most closely approximated in sequences of fossiliferous marine sediments. A land bridge existed in early Pliocene time; Hopkins (1967a, p. 454) stated "there is a clear record of repeated dispersals of land animals across Beringia through much of the Pliocene, during the interval from 10 to 4 million years ago." However, in late Pliocene time, about 3.5–4 m.y. ago, Bering Strait was reopened and permitted marine animal migrations. This is inferred from the presence of identical species of diagnostic Pliocene mollusks in both the North Pacific and the North Atlantic at this time (Durham and MacNeil, 1967). Late Pliocene sediments interbedded with basaltic lava flows and tillites on the Tjornes Peninsula of Iceland record an influx of Pacific mollusk species, reflecting the opening of Bering Strait about 3.5 to 4 million years ago (Einarsson, 1967) (table 1). It also now seems possible to correlate certain marine deposits in Alaska with the Pliocene Coralline Crag of England and certain other Alaskan marine deposits with the Pleistocene Red Crag of England on the basis of the common occurrence of several members of rapidly evolving stocks of *Neptunea*, *Chlamys*, and *Astarte*. Most of the beds diagnosed as Pliocene in Alaska on the basis of these genera also contain mollusks apparently restricted to Pliocene beds on the Californian, Siberian, and Japanese coasts. Alaskan beds diagnosed as Pleistocene contain few or no forms indicative of Pliocene age elsewhere on North Pacific shores.

One of the most complete records of late Cenozoic glacial and faunal data in the world is that reported from the Gulf of Alaska. It is also here that a good chance exists for the accurate placement of the Pliocene-Pleistocene boundary. The Yakataga Formation has long been known to contain Pliocene and Pleistocene fossiliferous marine tillites (Taliaferro, 1932; Miller, 1953a, 1957; MacNeil, 1967; Bandy and others, 1969). The most critical exposure is the well-documented section of 1,220 m of conglomerate, siltstone, and conglomeratic sandy mudstone beds on Middleton Island (pl. 1; Miller, 1953a, table 1).

On the basis of the study of pectinids in the section, MacNeil (1967, p. 32) suggested that the Pliocene-Pleistocene boundary be placed at the top of unit 17 of Miller's section. Unit 17 is about 300 m from the top of the section and marks the position where *Chlamys pseudislandica plafkeri* first appears. The top of the range of *Chlamys (Leochlamys) tugidakensis* lies immediately below and is Pliocene (table 1).

Because of the critical nature of the Middleton Island section in regard to the Pliocene-Pleistocene boundary, Plafker, Mankinen, and Doell (written commun., May 9, 1972) attempted to relate the section to the standard paleomagnetic scale. On the basis of preliminary examination, they inferred that the section covers the time interval from about 2 to 0.7 m.y. ago (table 1). Unit 17 would be at about 1.5 m.y. ago, if the deposition rate were constant.

Subsequent to MacNeil's (1967) report, additional fossil collections have been made, and Plafker stated (oral commun., May 11, 1972) that *C. plafkeri* now is more widespread and occurs lower in the section than previously reported. Future studies will undoubtedly reframe the placement of Pliocene-Pleistocene boundary on Middleton Island.

Small late Pliocene or early Quaternary floras (of pollen and wood) provide broad correlations between marine and nonmarine sediments in Alaska. However, an essentially modern taiga flora (spruce to birch) has been found in sediments underlying a 5.7-m.y.-old lava flow 30 km south of Cape Deceit (Hopkins and others, 1971). More details will be needed to position the Pliocene and Pleistocene boundary on the basis of floras.

Paleomagnetic and potassium-argon studies made on some of the widespread and almost continuous late Tertiary and Quaternary volcanic sequences in Alaska aid in relating the events near the Pliocene and Pleistocene boundary to the worldwide geomagnetic chronology. A rather complete record of late Cenozoic volcanism is documented by paleomagnetic and potassium-argon studies on Nunivak Island (Hoare and others, 1968) (table 1).

GLACIAL GEOLOGY

Because of the difficulty of correlating one isolated area in Alaska with another and the early impossibility of relating Alaskan glacial sequences to the better known glacial sequences of the north-central United States and western Europe, provincial names were given to many local Alaskan glacial chronologies established during the late 1940's and the 1950's. As field and photogeologic studies have progressed, however, it has become possible to trace or compare sequences of moraines in series of adjoining valleys throughout large areas. Radiometric dates, significant stratigraphic relations with marine sediments and other interglacial deposits, and an improved understanding of the nature and rates of weathering and denudation have now made it possible to make correlations with considerable confidence within single mountain ranges and with somewhat less confidence between different mountain ranges (table 2). It has also become possible in a general way to separate drift sheets in most areas into units of pre-Illinoian, Illinoian, Wisconsinan, and post-Wisconsinan ages. By no means have correlation problems been completely eliminated; nevertheless, progress has been sufficient to permit the preparation of a glacial map of Alaska at a scale of 1:2,500,000 (Coulter and others, 1965); figure 6 is based upon this compilation but incorporates certain modifications resulting from new information since the map was compiled and from the writer's differences of opinion concerning certain correlations.

Eventually, it will no doubt be possible to relate pre-Illinoian Quaternary events in Alaska to the classical glacial sequences of north-central United States and western Europe. However, until the earlier glacial sequences in the classical areas, as well as in Alaska, are dated accurately, terms such as Nebraskan and Kansan probably should not be applied to pre-Illinoian glacial events in Alaska, and only provincial names should be used (table 2).

The most reliable correlations are based on the tracing of moraines from one area to another, on radiocarbon dating, and on similar stratigraphic relations. A comparison of the topographic expression of moraines of various ages in widely separated areas is more difficult and less reliable because, in Alaska, denudational processes differ in nature and rate of operation from one region to another. Moraines are modified much more rapidly by solifluction and frost creep in northern Alaska than in southern Alaska; stream erosion and gullying probably progress more rapidly in southern Alaska than in northern Alaska. Furthermore, the surfaces of the older moraines differ considerably, even within small areas, as a result of local and regional differences in the thickness of the loess mantle.

Glaciers have covered about 50 percent of the present area of Alaska at one time or another, but large areas in central and northern Alaska have never been glaciated (fig. 6). Deposits assigned to each of the four major Pleistocene glaciations recognized in north-central United States are present in at least some parts of Alaska. The extents of the late Tertiary and early Pleistocene glacial advances remain almost unknown, but the extents of middle and late Pleistocene and Holocene advances are well established.

During Illinoian, Wisconsinan, and Holocene time, glaciers were much more extensive in southern Alaska than in northern Alaska and much more extensive on the south flanks than on the north flanks of individual mountain ranges, indicating that the glaciers were nourished chiefly by air masses moving northward or northeastward from the northern Pacific Ocean. In western Alaska glacial ice was much more extensive during Illinoian time than during Wisconsinan time, but in eastern Alaska the maximum Illinoian and Wisconsinan glacial advances differed very little. Illinoian snowline was about 170 m lower than Wisconsinan snowline throughout Alaska, and the greater difference between the extent of Illinoian and Wisconsinan glaciers in western Alaska seems to reflect the wide distribution of uplands there that were just high enough to serve as source areas for glaciers during Illinoian time but not high enough to serve as source areas during Wisconsinan time.

Wisconsinan time as used in this report encompasses the interval that started at the end of the Sangamon Interglaciation (approximately 70,000 years ago) and ended about 10,000 years ago. The maximum Wisconsinan ice advance was attained during early Wisconsinan time; advances of late Wisconsinan time were less extensive. Glacial advances during the past

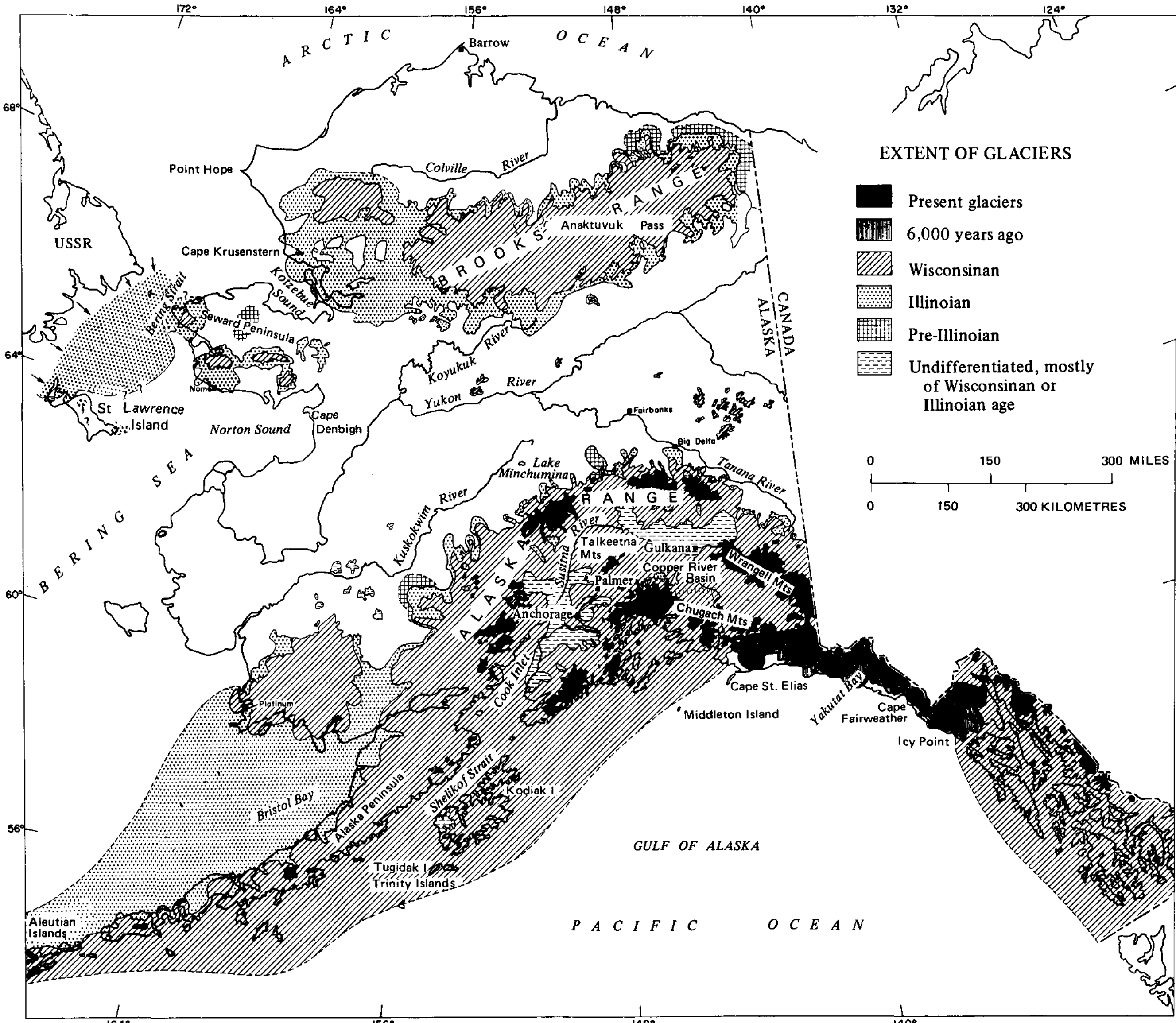


FIGURE 6.—Extent of Quaternary glaciations in Alaska. Modified from Coulter, Hopkins, Karlstrom, Péwé, Wahrhaftig, and Williams (1965). Additional information on Seward Peninsula and Bering Strait from D. M. Hopkins and C. L. Sainsbury (written commun., 1968) and Nelson and Hopkins (1972).

10,000 years were much less extensive than any late Wisconsinan glacial advance in most parts of Alaska, but Holocene glacial advances that culminated 1,000–2,000 years ago extended well beyond any recognized glacial advance of Wisconsinan age on the Pacific coastal plain and the Pacific slope of the Chugach Mountains between Cape St. Elias and Russell Fiord (the eastern arm of Yakutat Bay) (Plafker and Miller, 1958; D. J. Miller, oral commun., 1961). The anomalously large extent of Holocene glaciation between Cape St. Elias and Russell Fiord may reflect great uplift of this tectonically active coastal area during late Wisconsinan and Holocene time. Reid (1970a, b) showed that the late Wisconsinan terminal moraine of Martin Glacier, a few miles west of Cape St. Elias, marks the most extensive advance of that glacier on land; earlier it was a tidewater glacier. Reid concluded that this phenomenon too is probably the result of the tectonically active coast during late Wisconsinan and Holocene time.

EVIDENCE OF MIOCENE TO EARLY PLEISTOCENE GLACIATION

The strongly deformed Yakataga Formation and correlative deformed sedimentary sequences exposed between Cape Spencer and Cape St. Elias, on Middleton

Island, and on some of the Trinity Islands south of Kodiak Island contain many thick beds of conglomeratic sandy mudstone. This mudstone contains abundant striated clasts; its completely ungraded texture and the abundant mollusk shells in apparent growth positions indicate that the sediment consists of material "rained" upon the sea bottom rather than material transported laterally by turbidity currents or submarine landslides (Miller, 1953a, 1957). For these reasons, this mudstone is interpreted as glacial drift that accumulated either beneath an ice shelf or beneath icebergs broken from tidewater glaciers in Tertiary time (table 1). The composite outcrop thickness of the Yakataga Formation is about 5,500 m (Plafker, 1967, 1971). The base of the Yakataga Formation is of early to middle Miocene age on the basis of its molluscan fauna (Miller, 1961b, fig. 46; Plafker, 1971). Foraminifera, however, indicate that it could be as young as late Miocene (Bandy and others, 1969). Cooling is first indicated by the influx of a cold-water molluscan fauna (Addicott and others, 1970, p. 5c) and the cold-water foraminifera *Turborotalia pachyderma* at about the same stratigraphic horizon as the lowest beds containing marine glacial deposits. These faunas indicate decreases of surface water temperatures of 10°–15°C in early Yakataga time. The youngest beds of the Yakataga Formation, which are exposed at the top of a 1,170-m-thick section on Middleton Island, are presumably of early Pleistocene age according to F. S. MacNeil (in Plafker, 1967, 1971). At least one sheet of tillite at Middleton Island, in the upper part of the section was deposited by grounded glaciers, not by icebergs or shelf ice (D. M. Hopkins, written commun., 1968).

In an area in the eastern Wrangell Mountains, Capps (1915a, 1916b, p. 63–67) reported at least 10 layers of supposed tillite in a deformed sedimentary sequence that is clearly no younger than early Pleistocene in age, but it seems possible that the tillite layers actually represent volcanic agglomerate or volcanic mudflow deposits. Denton and Armstrong (1969) demonstrated that these deposits are tillites, and that the earliest tillite is about 10 m.y. old. In the Arctic Ocean, work on magnetic stratigraphy, coupled with faunal patterns in deep sea cores taken from Ice Island T–3, suggests that glaciation began more than 4 m.y. ago (Steuerwald and others, 1968; Clark, 1969, 1971). Herman (1970; Herman and others, 1971) suggested that glaciation began 2.4 m.y. ago (table 1).

Thus, D. J. Miller's pioneering work and the subsequent work of others in southern Alaska indicate that glaciation in Alaska started at least in late Miocene time, and perhaps as early as early middle Miocene time. This conclusion agrees with the data suggesting late Tertiary glaciation in Antarctica (table 1) (Denton and others, 1969, 1971) and Iceland (Einarsson, 1967).

EARLY PLEISTOCENE GLACIATIONS

Local patches of pre-Illinoian drift have been recognized in most of the glaciated areas in Alaska; widespread deposits are found only along the coast of the Gulf of Alaska, on Middleton Island, possibly in the upper part of the Yakataga Formation, and in a few places on the north flanks of the Alaska Range and the Brooks Range. Many areas are too small to be shown in figure 6. Throughout much of Alaska, glacial advances of Illinoian age evidently were approximately coextensive with, or more extensive than, the early Pleistocene advances. Where early Pleistocene drift is present, it is found only in isolated patches on ridges above the limits of ice advances during Illinoian time, in rare exposures beneath Illinoian drift, or in areas slightly beyond the outer limits of Illinoian drift.

Along the Gulf of Alaska coast, drift of early Pleistocene age occurs in areas where ice advances of Holocene age extended well beyond the limits of late Pleistocene ice advances. Here, drift is assigned pre-Illinoian age on the basis of its degree of deformation and induration and of molluscan faunas enclosed in intercalated marine sediments. The fauna suggests that the beds are equivalent in age to the Beringian, Anvilian, and Einahnuhtan transgressions (Hopkins, 1967b, p. 57). Two large shells from Pleistocene beds on Middleton Island have been dated by the U^{238}/Th^{230} method as being 210,000 ± 20,000 years and 220,000 ± 20,000 years (Blanchard, 1963), which places the beds in late Pleistocene time. However, an extensive molluscan fauna collected from these same beds indicates to F. S. MacNeil an early or early middle Pleistocene age (in Plafker, 1971). Furthermore, a complex postdepositional tectonic history—involving northwestward tilting at an average angle of 28°, truncation of the tilted strata, uplift above sea level, and offset by active minor faults—suggests to George Plafker (written commun., Apr. 21, 1970) that the radiometric dates are probably too young.

In areas away from the Gulf of Alaska coast, the early Pleistocene drift typically consists of patches of till or sprinklings of erratic boulders scattered over old erosion surfaces high above the present valley bottoms. There are no recognizable glacial landforms, such as end moraines, but in the north-central Alaska Range, Wahrhaftig (1958, p.22–27) distinguished the outer limit of his Browne Glaciation of early Pleistocene age by concentrations of boulders on a loop pattern enclosing the lower middle course of the lower Nenana River.

Karlstrom (1964, pl. 1) mapped extensive areas covered by pre-Illinoian glaciers on the north side of the Alaskan Range from the Tonzona River to the Delta River. Many glacial geologists working in this area do not hold this view, and this controversy gave rise to the straight-line boundary of the early Pleistocene extent of glacier ice along the 153 degree meridian on the map of glacial extents in Alaska (Coulter and others, 1965). The writer believes the glaciers of early Pleistocene time were not as extensive as shown by Karlstrom. One point to consider is that these early glaciations may have occurred 1–2 m.y. ago or earlier, and the Alaska Range may not have been as high as it is now. This controversy may stimulate further work on the north side of the Alaska Range.

Submarine drift and glacial topography are now known on the sea floor in the Bering Strait (fig. 6) (Grim and MacManus, 1970; D. M. Hopkins, oral commun., 1969) and are thought at this time to represent both Illinoian and pre-Illinoian glacier advances or perhaps an icecap in the Strait area. Carsola (1954) indicated a lack of glaciation of the Continental Shelf of the Chukchi Sea to the north.

Precise age assignments can be given to a few of the occurrences of drift of early Pleistocene age. Hopkins (in Hopkins and others, 1960, p.61) considered the single sheet of pre-Illinoian drift in the coastal plain at Nome to be early and middle Pleistocene because it underlies the marine sediments of "Intermediate Beach." "Intermediate Beach" was formerly considered to be of Yarmouth age, but it is now (Hopkins, 1967a) assigned to the Anvilian transgression because its molluscan fauna contains many more extinct species and in other respects contrasts sharply with the Kotzebuan fauna in marine sediments of later pre-Illinoian (Yarmouth?) age in the Kotzebue Sound area (Hopkins and others, 1962; McCulloch and others, 1965; Hopkins, 1967a). The writer tentatively considers the Browne and Dry Creek Glaciations in the north-central Alaska Range to be two early and middle Pleistocene advances because they were separated from one another and from the later Healy Glaciation of probably Illinoian age by long periods during which considerable regional tilting and stream erosion took place (Wahrhaftig, 1958, p. 31, 41–42). Recently, Clyde Wahrhaftig (oral commun., July 4, 1970) suggested that the Browne Glaciation is more than 2.7 m.y. old on the basis of field relations and radiometric dating near Jumbo Dome on the north side of the Alaska Range and therefore is pre-Pleistocene in age.

Karlstrom (1960a) considered the Mount Susitna and Caribou Hills Glaciations of Cook Inlet areas to represent the Nebraskan and Kansan Glaciations, but without dates all that can be said is that they are pre-Illinoian in age. Glacial landforms assigned to these stages are much more modified than those of the Eklutna Glaciation, and the Illinoian age of the Eklutna Glaciation seems well established.

GLACIATION OF ILLINOIAN AGE

DISTRIBUTION OF DRIFT

Prior to 1945, where abundant erratic boulders forced the recognition of a glacial event, the glaciation was ascribed to an ice advance of Wisconsinan age. Studies since 1945 have shown, however, that drift of Illinoian age is present in nearly every glaciated area in Alaska. There is general agreement as to which areas of surface drift should be assigned to the Illinoian glacial cycle throughout most of Alaska. However, uncertainties persist for areas in southeastern Alaska and on the north flank of the Brooks Range.

The largest areas of Illinoian drift exposed at the surface are in western Alaska in the Kotzebue Sound and Bristol Bay areas and on Seward Peninsula (fig. 6). Farther eastward along the flanks of the Brooks Range and along the north flank of the Alaska Range, Illinoian end moraines lie only short distances in front of moraines of Wisconsinan age. An Illinoian glaciation has been identified on the Pribilof Islands (Hopkins and Einarsson, 1966). No till of either Illinoian or Wisconsinan age is recognized along the Alaskan coast between Cape Fairweather and Russell Fiord; apparently glaciers in this area were thinner and less extensive during Illinoian and Wisconsinan time than they have been during the last few thousand years. Little attention has been given to the pre-Holocene glacial history of southeastern Alaska, and no attempt has been made there to recognize traces of glaciation of Illinoian age.

Distribution of drift of Illinoian age in far western Alaska has been clarified recently. Sainsbury (1967a, b; oral commun., 1969) stated that Skull Creek erratics are widespread in western Seward Peninsula, and he believed that much of the Bering Strait was covered by ice at this time. One of the active areas of glacial research in Alaska today centers on the extent and age of glaciations in Bering Strait and adjacent Bering and Chukchi Seas and the relation to sea level. As indicated in figure 6, glaciers from the Chukotka Peninsula of U.S.S.R. adjacent to the Bering Strait once extended southward and eastward more than 100 km onto the Bering shelf, reaching St. Lawrence Island (Hopkins and others, 1972, p. 125). Hopkins (1972) believed that the glaciers could hardly have extended this far if it were a time of high sea level and glacial margins were afloat as suggested by Soviet workers. Grim and McManus (1970) indicated large-scale deformation structures on the sea floor probably caused by glacial ice in contact with the ground surface. Hopkins (1972) further concluded that this extent of Siberian ice then occurred during Illinoian time rather than during a time of high sea level, the Kotzebuan transgression.

New work by Hopkins has extended the limits of the

Nome River Glaciation in the southern part of the Seward Peninsula and in Bering Strait (compare fig. 6 this paper and fig. 3 in Péwé, Hopkins, and Giddings, 1965).

BASIS FOR AGE ASSIGNMENT

Drift of Illinoian age is most firmly identified on stratigraphic and radiometric evidence near Nome and in the shore bluffs of Kotzebue Sound and Cook Inlet. Elsewhere, drift sheets are assigned to the Illinoian Glaciation on the basis of similarities in surface expression and in positions in local glacial sequences to the Illinoian drift in the Nome, Kotzebue Sound, and Cook Inlet areas. The drift of Illinoian age at Nome (till and outwash of the Nome River Glaciation—Hopkins and others 1960, p.51) overlies marine sediments of an earlier Pleistocene interglacial interval and is covered locally by marine sediments of Sangamon age from which shells have been dated by radiocarbon as more than 38,000 years old (W–810, Rubin and Alexander, 1960, p. 174) and by the U_{238}/Th_{230} method as 100,000 years old (Blanchard, 1963). The Illinoian drift on the shores of Kotzebue Sound was first described by Hershey (1909). More recent work (Hopkins and others, 1962; McCulloch and others, 1965) showed that the till overlies marine sediments dated by the U_{238}/Th_{230} method as 175,000±16,000 and 170,000±17,000 years (Blanchard, 1963); this drift is covered in some places by marine sediments and in others by a forest bed more than 38,000 years old (W–344, Rubin and Alexander, 1958, p. 1483) of probable Sangamon age. Here sediments of Sangamon age are overlain in turn by unoxidized loess of probable Wisconsinan age.

In the Cook Inlet area near Anchorage, oxidized drift of the Eklutna Glaciation of Illinoian age is overlain by unweathered drift of the Knik Glaciation of early Wisconsinan age (Miller and Dobrovolny, 1959, p. 11; Karlstrom, 1964, p. 33; Trainer and Waller, 1965). On St. George Island in the Pribilofs (fig.1), Illinoian till is bracketed by marine deposits of Kotzebuan (pre-Illinoian) and Pelukian (Sangamon) age (Hopkins and Einarsson, 1966).

EVIDENCE OF MULTIPLE GLACIAL ADVANCES

Most areas of Illinoian drift contain more than one recognizable morainal ridge, suggesting that the Illinoian glacial cycle, like the Wisconsinan cycle, was a complex event consisting of several glacial advances, retreats, and readvances. Stratigraphic evidence of multiple glacial advances during Illinoian time has been found, however, only in the Selawik Lake area, inland from the southeast corner of Kotzebue Sound. Here, D. M. Hopkins, D. S. McCulloch, and R. J. Janda (unpub. data, 1961) found two exposures about 40 km apart in which marine and fluvial sediments about 175,000 years old are overlain by two sheets of Illinoian drift separated from one another by a few metres of stratified silt and peat.

SURFACE MORPHOLOGY

The degree of preservation of primary microrelief on the Illinoian drift varies greatly in different physiographic settings. The larger primary relief features resulting from glacial erosion and sedimentation, cirques, U-shaped valleys, major morainal ridges, and outwash terraces, remain recognizable, but smaller features, such as minor morainal ridges, kames, eskers, and kettles have been modified by and commonly obscured by mass wasting, deposition of colluvium from nearby slopes, and dissection by consequent gullies. In many places, the addition of a mantle of loess, several metres thick, of late Illinoian and Wisconsinan age has contributed further to the obliteration of minor surface features on the Illinoian moraines.

Illinoian moraines consist of large, smoothly rounded ridges with irregular crests and gentle slopes that are devoid of primary microrelief features (fig. 7) and that are incised by consequent gullies at intervals of about 1 km. In the northeastern part of the Alaska Range, Péwé and Holmes (1964) and Holmes and Foster (1968) showed that slopes are consistently more gentle on moraines of the Delta (Illinoian) Glaciation than on moraines of the Donnelly (Wisconsinan) Glaciation. Lakes are common on Illinoian moraines, but nearly all are "second cycle" thaw lakes developed in initial low areas on the surface of the drift. The Delta moraine of Illinoian age contains only 1.2 kettle ponds per km^2 while the adjoining Donnelly moraine of Wisconsinan age contains an average of 6.2 kettle ponds per km^2; bogs, however, are twice as abundant on the Delta moraine as on the Donnelly moraine (Péwé and Holmes, 1964; Holmes and Benninghoff, 1957, p. 63–69). Mining excavations in Illinoian drift near Nome show that original kettle holes have been obliterated by a fill of colluvial silty gravel several metres deep (D. M. Hopkins, unpub. data, 1955–1957). Cirques that were occupied by Illinoian ice but have remained vacant ever since display the typical hemiamphitheater form, but the original rock cliffs at the head walls have been reduced to lines of low rock bluffs 2–6 m high on the upper slopes, adjoined by long smooth aprons of colluvium or vegetated talus.

SOILS AND WEATHERING PROFILES

Drift of Illinoian age commonly displays a soil and weathering profile several times deeper and much more intensely developed than profiles of Wisconsinan drift in the same area. The post-Illinoian weathering profile is most easily interpreted as a product of surficial weathering processes in places where it is developed in fine-grained sediments such as till or glaciolacustrine

FIGURE 7.—High oblique aerial photograph of part of a terminal moraine of Illinoian age along the east side of Delta Creek, interior Alaska. View looking southwest toward Alaska Range. Photograph by U.S. Army Air Corps, April 24, 1947, 10,000 feet elevation. Photograph No. 75RT-46PL-m-7m56-46RS-24 Apr 47-65A.

silt and clay; well-exposed fine-grained sediments of Illinoian age are easily distinguished from similar sediments of Wisconsinan age on the basis of their weathering profiles. In coarse sediments such as outwash sand and gravel, deep and intense oxidation caused by surficial weathering frequently has been confused with discolorations resulting from the redistribution of iron by ground water (for example, see discussion in Miller and Dobrovolny, 1959, p. 33–34); thus, distinctions between coarse sediments of supposed Illinoian and Wisconsinan age based only upon the presence or absence of oxidation colors must be regarded with suspicion.

The post-Illinoian weathering profile on Illinoian drift is deeper and better developed in southern Alaska than in northern Alaska. Till, silt, sand, and gravel of the Eklutna (Illinoian) Glaciation near Anchorage are oxidized to depths of 10 m or more (Miller and Dobrovolny, 1959, p.11; Karlstrom, 1964). Drift of Illinoian age in mining excavations near Platinum is oxidized to depths of nearly 15 m (D. M. Hopkins, unpub. data, 1957). Illinoian till near Nome and on the shores of Kotzebue Sound shows oxidation to depths of about 3 m, and an excavation in an Illinoian esker showed a well-defined Arctic Brown soil profile more than 1.7 m thick (D. M. Hopkins, D. S. McCulloch, and R. J. Janda, unpub. data, 1961). Till of the Delta (Illinoian) Glaciation in the northeastern Alaska Range is oxidized to depths of 1.2–3 m near Big Delta (Holmes and Benninghoff, 1957, p. 79; Holmes and Péwé, 1965) and to 3 m near the Johnson River (Holmes, 1965, p. H10; Holmes and Foster, 1968). No descriptions of weathering phenomena in drift of Illinoian age are available for the north flank of the Brooks Range.

Some investigators (for example, Péwé and others, 1953; Péwé and Holmes, 1964; Holmes and Benninghoff, 1957, p. 79; Wahrhaftig, 1958, p. 32) found a high proportion of rotten clasts, especially of granitic rocks, in near-surface zones in the Illinoian drift. In western Alaska, however, D. M. Hopkins (unpub. data, 1957) was unable to recognize a consistent relation between abundance of weathered clasts and age of the enclosing drift.

A deep post-Illinoian weathering profile underlies drift of early Wisconsinan age in several places in the Cook Inlet area (Miller and Dobrovolny, 1959, p. 11). The post-Illinoian weathering profile is covered by loess of Wisconsinan age on the shores of Kotzebue Sound (Hopkins and others, 1962; McCulloch and others, 1965), indicating that most of the weathering was accomplished during the Sangamon Interglaciation.

SNOWLINE

The writer and his coworkers made a series of compilations of the former positions of snowline during Illinoian and Wisconsinan time as interpreted from the altitudes of cirque floors in various parts of Alaska. Illinoian snowline lay approximately 500–600 m below present snowline and about 150–250 m below Wisconsinan snowline throughout a large part of Alaska (figs. 8, 9). Snowline was highest in the dry continental climate of east-central Alaska and lay at an altitude of about 1,250 m during Illinoian time (Péwé and Burbank, 1960; Péwé and others, 1967). It sloped downward to an altitude of less than 300 m on Seward Peninsula in western Alaska (fig. 8). Snowline lay appreciably lower on the south sides than on the north sides of individual mountain ranges.

Hopkins (written commun., 1968) suggested that at one time the Illinoian(?) snowline intersected present sea level on the Aleutian Ridge because glacial ice came from present water areas southwest of Amchitka Island (pl.1) and overrode the east end of the island. On St. George Island 700 km northwest of the Aleutians, the Illinoian snowline was at an altitude of 150 m (Hopkins and Einarsson, 1966).

SIGNIFICANCE OF THE DISTRIBUTION PATTERN OF ILLINOIAN DRIFT

In Western Alaska, glaciers were vastly more extensive during Illinoian time than during Wisconsinan time, but the difference in extent decreases eastward. However, snowline during Illinoian time lay about 170 m below the position of snowline during Wisconsinan time (fig 8) (Péwé and others, 1967) throughout a large part of Alaska. The relatively greater extent of glaciers during Illinoian time in western Alaska evidently is a result of the wide distribution of upland areas that were just high enough to support cirques and small icecaps during Illinoian time and that were not high enough to project above snowline during Wisconsinan time.

The pronounced asymmetry in the extent of glaciation and the position of snowline on all mountain ranges of Alaska, including the Brooks Range, indicate that during both the Illinoian and Wisconsinan glacial cycles, the glaciers were nourished primarily by moisture from airmasses arriving from the south or southwest. This feature was first pointed out by Capps (1912) and restated by Karlstrom (1964) and later workers. Neither the common features in the two patterns of glaciation nor their differences seem explicable in terms of changes in the amount of moisture that might have been contributed by airmasses entering Alaska from the Arctic Basin as suggested by Ewing and Donn (1956, 1958). The Bering shelf, constituting approximately the northwestern third of the Bering Sea, was evidently dry land during most of Wisconsinan time and probably during most of Illinoian time (Hopkins, 1959b). Airmasses moving into Alaska during these glacial intervals must have obtained most of their moisture while passing over the north Pacific Ocean and western Bering Sea.

GLACIATION OF WISCONSINAN AGE

DISTRIBUTION AND EXTENT

Drift of Wisconsinan age is found in nearly all the mountainous areas of Alaska (fig. 6). North of the crest of the Alaska Range and of the Alaska Peninsula, the glaciers were essentially alpine in character; they filled mountain valleys and in places spread as piedmont lobes in adjoining lowlands. However, most lowland and upland areas in central and northern Alaska were free of ice. Areas south of the crest of the Alaska Range and the Alaska Peninsula were almost completely inundated by ice, and icecap conditions prevailed over large areas. Glaciers originating in the mountains filled adjoining lowland basins such as the broad Susitna valley and spread onto the then-dry continental shelf, thickening to form icecaps as much as several tens of kilometres wide and several hundreds of kilometres long. An icecap in Shelikof Strait sent outlet glaciers northward through low passes in the mountainous backbone of the Alaska Peninsula (pl.1) to spread as piedmont lobes on the coastal plain along the southeastern coast of Bristol Bay; other ice tongues pushed from Shelikof Strait southward into valleys in southwestern Kodiak Island

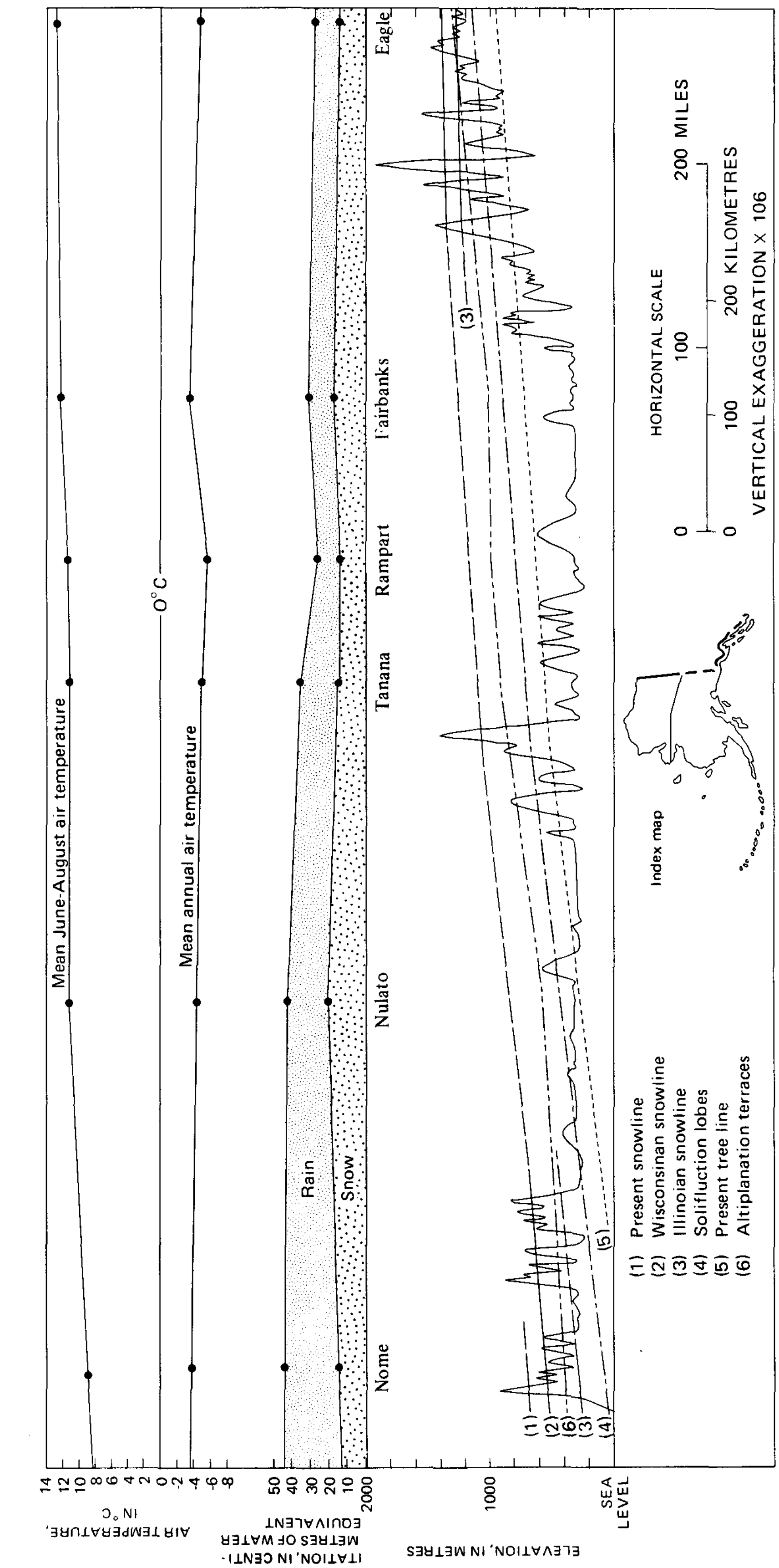


FIGURE 8.—East-west section across central Alaska illustrating change of certain meterological parameters and elevations of present snowline, snowlines of Wisconsinan and Illinoian age, well-developed altiplanation terraces, solifluction lobes, and present tree line. Compiled by T. L. Péwé and L. R. Mayo. (Past snowlines were determined from the base of cirques.)

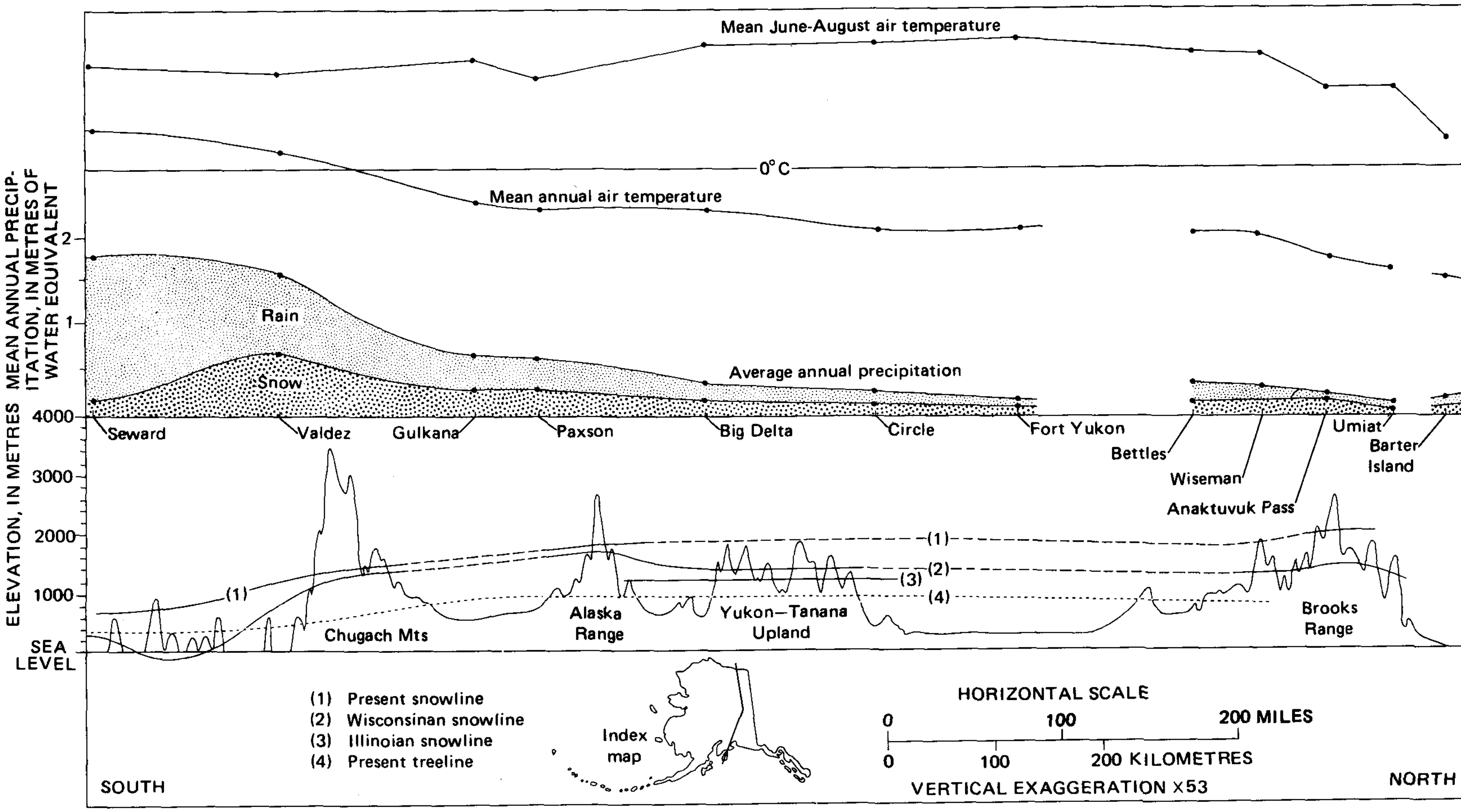


FIGURE 9.—North-south section across central Alaska illustrating change of certain meterological parameters and elevations of present snowline, snowlines of Wisconsinan and Illinoian age, and present tree line. Climatological data from the Brooks Range is modified from Porter (1966). Those stations are about 250 km west of the section. Compiled by T. L. Péwé and L. R. Mayo. (Past snowlines were determined from the base of cirques.)

that were too low to support local alpine glaciers during the Wisconsinan glacial cycle (Coulter and others, 1965; E. H. Muller, Clyde Wahrhaftig, and T. N. V. Karlstrom, oral commun., 1955–57). Similar icecaps accumulated in several places on the southern half of the Aleutian Ridge and moved northward over the lower parts of the present islands (Judson, 1946; Bradley, 1948; Scruton, 1953; Coats, 1956; Byers, 1959; Powers and others, 1960; Hopkins and Einarsson, 1966). Wisconsinan drift is lacking along the coast of the Gulf of Alaska between Cape St. Elias and Russell Fiord. Glaciers there may have been less extensive during Wisconsinan time than during Holocene time; apparently mountains north of this part of the Gulf of Alaska coast were lower during Wisconsinan time than at present and consequently supported less extensive glaciers (D. J. Miller, oral commun., 1960).

Wisconsinan glaciers pushed south into Prince William Sound and are in part responsible for the present submarine topography (Von Huene and others, 1967). The increase in submarine geological work in Prince William Sound and adjacent areas, stimulated by the 1964 Alaskan earthquake, has already enlarged our views of glacial action in this area and may lead to a better understanding of ice shelf deposits.

Deglaciation during late Wisconsinan time in areas of alpine glaciation north of the crest of the Alaska Range seems to have consisted of a gradual shrinkage in the thickness and extent of ice tongues that remained active even in their frontal areas. However, deglaciation in areas of large coalescent and piedmont ice masses south of the Alaska Range was much less orderly. Widespread dead-ice features, such as areas of kame-kettle topography, networks of eskers, and successions of (annual?) marginal channels in the wider valleys and lowland areas, record ice stagnation and down-wastage without further movement in large areas. Dead-ice features are especially abundant in the lower Matanuska Valley northeast of the head of Cook Inlet (Trainer, 1960), in parts of the Copper River Basin (D. R. Nichols and J. R. Williams, oral commun., 1955–60), and in the belt of the lowlands just south of the Alaska Range between the Alaska Railroad and the Richardson Highway (Péwé, 1961c, 1965c; Kachadoorian and others, 1954; Jahn, 1966, fig. 38, p. 127).

BASIS FOR AGE ASSIGNMENT

The Wisconsinan glacial cycle in Alaska was clearly a complex event consisting of at least two major glacial advances; several minor oscillations of ice fronts occurred during the latter of the two major glacial events (table 2). In several areas the two major divisions of the Wisconsinan glacial cycle are well established by stratigraphic investigations and radiometric dating as approximately equivalent to the early and late Wisconsinan glacial cycles in north-central United States. Some of these areas are Cook Inlet (Karlstrom, 1964; Trainer and Waller, 1965), southwestern Brooks Range (Hamilton, 1969), Chagvan Bay (Porter, 1967), and York Mountains (Sainsbury, 1967a, b) (table 2).

Records of early Wisconsinan Glaciation have not yet been recognized in some areas. Sainsbury (1967a, b) pointed out that early Wisconsinan deposits have not yet been mapped in the Kigluaik Mountains of the Seward Peninsula but that they probably do exist.

Drift of Wisconsinan age in the shore bluffs of Cook Inlet has been dated as a result of detailed studies by Miller and Dobrovolny (1959) and by Karlstrom (1965). An abundance of stratigraphic observations and radiometrically dated organic specimens establish that the Knik Glaciation is of post-Sangamon age but older than 37,000±2,000 years, and thus it is of early Wisconsinan age (Karlstrom, 1964, p. 57).

Karlstrom (1957, 1960a) considered the Knik drift to be of "post-Illinoian, pre-Wisconsinan (Sangamon)" age, and Miller and Dobrovolny (1959, p. 15) considered it to be pre-Wisconsinan but not necessarily as old as Sangamon or Illinoian. These differing age assignments reflect different concepts of the duration of the Wisconsinan Glaciation, rather than different opinions as to the absolute age and correlation of the Knik Glaciation.

Drift of the Knik Glaciation rests on interglacial deposits of Sangamon age and is covered by the Bootlegger Cove Clay. The clay consists of three units and was regarded in the past as estuarine, lacustrine, or marine. Smith (in Hanson, 1965, p. A20) studied the Foraminifera and concluded that the clay is entirely marine. After deposition of the middle unit, the Naptowne glacial advance occurred in the Anchorage area.

During the fifties and sixties, the Bootlegger Cove Clay was thought to be middle Wisconsinan in age and indicative of a higher stand of sea level. A U_{238}/Th_{230} date of 38,000–48,000 years (Blanchard, 1963) on shells from the clay supported this concept. The Naptowne Glaciation was thought to be late Wisconsinan in age (Karlstrom, 1964).

Four recent radiocarbon age determinations and a new uranium series age on shells from the Bootlegger Cove Clay indicate that it was deposited about 14,000 years ago (Schmoll and others, 1972). The clay, designated the type deposit for the Woronzofian transgression, is now interpreted by Schmoll, Szabo, Rubin, and Dobrovolny (1972) as representing a marine transgression after the maximum development of the glaciation of Wisconsinan age, rather than an interstadial event of middle Wisconsinan age or an older glacial or interglacial event. The Naptowne advance is younger than 14,000 years and is not of conventional late Wisconsinan age as documented by Karlstrom.

In the Anaktuvuk Pass area of the central Brooks Range, Porter (1964) showed by radiocarbon dating that the Itkillik Glaciation of Detterman, Bowsher, and Dutro (1958) is late Wisconsinan in age instead of early Wisconsinan and also divided the Itkillik Glaciation into four stades (table 2), the youngest of which culminated about 6,200–8,300 years ago and the second oldest of which culminated about 13,300 years ago. On the south side of the central Brooks Range, Hamilton (1969) demonstrated by use of radiocarbon dating that his Siruk Glaciation is late Wisconsinan in age (younger than 31,000 years) and his Alatna Glaciation is early Wisconsinan in age (older than 38,000 years).

In the Chagvan Bay area southeast of the Kilbuck Mountains at Platinum (fig. 6), Porter (1967) reported multiple glaciations. Although early workers (Mertie, 1940; Hoare and Coonrad, 1961a, b) assigned the drift a Wisconsinan age, Porter recorded two glaciations of pre-Wisconsinan age and two of Wisconsinan. His Chagvan Glaciation is early Wisconsinan and more than 45,000 years old. The overlying Unaluk drift is termed late Wisconsinan and dated as older than 8,900 years but less than 45,000 years.

In the far western part of Seward Peninsula, Sainsbury (1967a, b) mapped drift of early Wisconsinan age overlying the Lost River terrace of Sangamon age and itself overlain by drift of a younger glaciation older than 10,000 years. Near Mount McKinley National Park organic material in the sediments of a former lake that had been partly confined by the outermost glacial ice at the culmination of the Riley Creek Glaciation in the Nenana River valley has been radiocarbon-dated as being 10,560±200 years old (W–49, Suess, 1954; also see Wahrhaftig, 1958, p. 45–46).

"Stop dates" establishing minimum ages of 8,000–12,000 years for Wisconsinan drift are available from the upper and lower Delta River areas (Péwé, 1965c, p. 66 and 91; Ives and others, 1964, p. 67–68) and in a few other areas, but throughout most of Alaska, the assignment of particular drift sheets to the Wisconsinan glacial cycle is based upon similarities in topographic expression and position in local glacial sequences to the Wisconsinan drift in dated positions in Alaska.

SURFACE MORPHOLOGY

Primary depositional and erosional topography on Wisconsinan drift is so sharply defined and well preserved that the glaciated character of the terrain is immediately obvious to the most casual observer (fig. 10). Mass wasting has modified the surfaces of moraines, valley walls, and cirques of the older advances of Wisconsinan age compared with those affected by the latest Wisconsinan advances, but the differences are subtle and emerge only upon detailed examination.

Frontal slopes on older Wisconsinan moraines appear to be slightly less steep than those on youngest Wisconsinan moraines, but no quantitative studies have been published demonstrating a consistent difference. Kettles on the older moraines commonly contain an appreciable thickness (as much as 3–5 m) of fine sediment and peat, and some kettles are now the sites of meadows and bogs; but, again, no studies demonstrating a systematic difference in the character of kettles on the older and younger Wisconsinan moraines have been reported. Older and younger Wisconsinan lateral moraines have not been observed to differ appreciably in most places.

The steep walls of alpine valleys and cirques glaciated during earlier and later Wisconsinan time differ more appreciably. The walls of valleys glaciated in earlier Wisconsinan time have been markedly regraded by mass wasting.

SOILS AND WEATHERING PROFILES

Soils and weathering profiles on drift of Wisconsinan age are much thinner and less conspicuous than those on drift of Illinoian age. Older Wisconsinan moraines commonly show soil profiles about twice as thick as those developed on the younger Wisconsinan moraines in the same area. In general, soils are thinner and less intensely developed as one proceeds northward or to higher elevations.

In the Anchorage area, till of the early Wisconsinan Knik Glaciation is oxidized to depths of 1–2 m where it is exposed at the surface and to depths of 35–60 cm where it is covered by the Bootlegger Cove Clay; no depths of oxidation are reported for till of the Naptowne Glaciation, but outwash of Naptowne age has a soil profile 60 cm deep in one exposure (Miller and Dobrovolny, 1959, p. 6, 22, 24). Till of late Wisconsinan age in the southwestern Copper River Basin is oxidized to a depth of 1 m (J. R. Williams, written commun., 1964).

Holmes and Benninghoff (1957, p. 84) reported soils only 30–90 cm deep, expressed by oxidation and secondary lime concentrations, in moraines of the Donnelly Glaciation probably of late Wisconsinan age. Donnelly till in the Johnson River area is weathered to a depth of 50 cm (Holmes and Foster, 1968, p. 32). Till of late Wisconsinan age at altitudes of 250–900 m along the Denali Highway on the south flank of the Alaska Range displays a well-developed podzol accompanied by oxidation extending to depths of 30–60 cm (Kachadoorian and others, 1954). D. M. Hopkins (unpub. data 1957) was unable to detect soil profiles on Wisconsinan drift on Seward Peninsula, but outwash and alluvial fan gravel of probable late Wisconsinan age show Arctic Brown soil profiles 30–60 cm thick. In the northeastern Brooks Range, Holmes and Lewis (1961) reported a "faint weathering profile approximately 1 m thick" and the presence of iron-stained and weathered rocks in the

FIGURE 10.—High-altitude oblique aerial photograph of part of a terminal moraine of Wisconsinan age along the Delta River in interior Alaska. The roughness of this moraine is typical of moraines of Wisconsinan age in Alaska. View looking southwest toward Alaska Range from over Delta Junction. Photograph by U.S. Army Air Corps, August 29, 1949, 10,000 feet elevation, photograph No. 206RT–55RT M864 55SRW 29 Aug. 49 9M58.

upper 60 cm of till of the Schrader Glaciation. Till of the Peters Glaciation lacks a perceptible weathering profile.

SNOWLINE

One of the parameters which yields quantitative information about the distribution of Pleistocene glaciers in Alaska as compared with today is the position of past and present snowlines. Such information is also valuable in speculating on Pleistocene climates. The term snowline is used in various ways, and also, different methods of calculations are in vogue. Definitions used here are those suggested by glaciologists and appear in the revised glossary of the American Geological Institute (1972) and in the International Hydrological Decade guide entitled "Seasonal Snow Cover" (UNESCO/IASH, 1970).

Climatic snowline. (1) The average line or altitude delineating, as of a specified time, the area with more than 50 percent snow cover on horizontal surfaces, averaged over a long time period (for example, 10, 30 years) of climatic significance. (2) The preceding line as observed in late summer so that it approximately coincides with the firn line or equilibrium line on glaciers. Cf.: regional snowline.

Equilibrium line. The level on a glacier where the net balance equals zero and accumulation equals ablation; the line separating the superimposed ice zone of the accumulation area (below). For some temperate mountain glaciers, it is very nearly coincident with the firn line, in which case it is common practice to use the latter term; on subpolar glaciers, the equilibrium line is lower than the firn line because freezing of meltwater occurs below the firn line forming superimposed ice. Synonym: equilibrium limit.

Firn line. (1) The highest level to which the winter snow cover retreats on a glacier. (2) The edge of the snow cover at the end of the summer season, thus the boundary between the superimposed ice zone below and the saturation zone above. Cf.: equilibrium line. Synonym: firn limit.

Glaciation limit. The lowest altitude in a given locality at which glaciers can develop or form, usually determined as below the minimum summit altitude of mountains on which glaciers occur but above the maximum summit altitude of mountains which have topography favorable for glaciers but on which no glaciers occur.

Orographic snowline. Term not recommended. See climatic snowline.

Regional snowline. The level above which, averaged over a large area, snow accumulation exceeds ablation year after year. Cf.: climatic snowline.

Snowline. (1) Momentary line delimiting an area or altitude with complete snow cover, or in a zone of patchy snow the area or altitude of more than 50 percent cover of snow. Synonym: transient snowline. Cf.: climatic snowline. (2) The line or altitude on land separating areas in which fallen snow disappears in summer from areas in which snow remains throughout the year; on glaciers it is identical with the firn line. Cf: orographic snowline, regional snowline, equilibrium line. Also written snow line.

MODERN SNOWLINE

Methods of Determination

Existing methods for determining the height of modern snowline was succinctly summarized by Østrem (1966, p. 126–129). The methods can be divided into three groups: (1) Those related to the firn line on glaciers, (2) those based on distribution of glaciers, and (3) those not directly connected with the study of glaciers.

Various methods have been used to plot the distribution of firn line, or climatic snowline of present usage. The best method is to make direct mass-balance measurements on a glacier for several years. From this it is possible to determine the average height of firn line (equilibrium line altitude, ELA); however, this method requires extensive fieldwork. A variation of this method is to use one or two glaciers, for which there are mass-balance measurements, as standards and extend the area of snowline distribution by using good topographic maps or aerial photographs taken at the end of the ablation season. This has been used in Norway (Liestøl, 1962) and western North America, including Alaska (Meier and Post, 1962).

Another technique for finding the position of equilibrium line altitude on glaciers is to use the accumulation area ratio (AAR). Accumulation area ratio is defined (Meier, 1962, p. 256) as the quotient of accumulation area divided by the total surface area of the glacier. Meier and Post (1962, p. 73) demonstrated that healthy glaciers (glaciers in dynamic equilibrium or advancing) have accumulation area ratios of 0.6 or more. Therefore by measuring the total surface area of the glacier, the equilibrium line is placed in that position on the glacier to permit an assumed accumulation area of that ratio. This method is gaining favor and has recently been used by Andrews and Miller (1972).

Other methods which are less accurate but entirely satisfactory require only good topographic maps and permit plotting of snowline over large areas. These methods are those in which climatic snowline (firn line) is (1) determined from an interpretation of the contour lines on a glacier to place the boundary between accumulation and ablation areas, (2) determined to be the mean elevation of the glacier, or (3) determined to be a line separating the accumulation and ablation areas at a position separating a percentage (one-quarter to three-quarters) of the lower part of the glacier from the upper part.

A method which eliminates considerable subjectivity, does not refer to firn line, and has wide use is the plotting of the "glaciation limit." Østrem (1964, 1966) showed that the altitude of glacier formation can be determined by studying the distribution of glaciers and the altitude of surrounding summits. It is possible to determine a critical height (the "glaciation limit") which has normally to be exceeded to form glaciers. Østrem (1966) stated that the glaciation limit is about 100 m higher than climatic snowline (firn line). This has been supported by work in Baffin Island (Andrews and Miller, 1972). The "glaciation limit" method works best where there are many small local glaciers; it is not applicable to icecap areas.

Another method for determining modern snowline disregards modern glaciers and relies instead on meteorological data. Modern snowlines are controlled in part by summer temperatures, and some writers report the elevation of modern climatic snowline may nearly coincide with the level of the mean 0°C summer or July isotherm (Leopold, 1951). By using the known lapse rate at a particular point and knowing the mean summer temperature near the ground, it is possible to calculate the elevation of the 0°C summer isotherm and thus, perhaps, elevation of modern climatic snowline at that particular locality. Charlesworth (1957, p. 9–10) concluded from his literature review, however, that the correlation is invalidated locally by the effects of precipitation and wind drifting. He stated (p. 9) that for climates between lat. 35° and 70° north, snowline coincides with the isotherm of 4°C±3° in the warmest month.

Modern Snowline in Alaska

The modern snowline in Alaska was probably first discussed by Romer (1929) and first sketched from glaciers on reconnaissance maps for eastern Alaska by Wasowicz (1929). A small-scale map of the equilibrium line altitude of the same area was plotted by Meier and Post (1962) from firn line and mass-balance measurements on five Alaskan glaciers and observations of firn lines on other glaciers from aerial photographs. A more detailed map of the equilibrium line altitude was prepared by Meier, Tangborn, Mayo, and Post (1971) for the entire western part of North America using the same techniques as in 1962. Karlstrom (1964) plotted modern snowline from cirque glaciers in south-central Alaska.

To test the technique for locating modern climatic snowline in Alaska by using meteorological data alone, several localities were analyzed where temperatures, lapse rates, and past snowline locations are known (Péwé and Reger, 1972). The climatic snowline level is close to 1,800 m elevation on two small north-facing cirques at the head of the east fork of Itikmalakpak Creek approximately 27 km east-northeast of Anaktuvuk Pass (pl. 1). Using the mean July temperature of Anaktuvuk Pass (10.3°C at 700 m elevation) and applying the lapse rate of 0.59°C/100 m suggested by Porter (1966, p. 87), one can extrapolate that the mean July temperature at the 1,800-m level is close to 6.8°C. A similar analysis with data provided by Larsson (1960) for Chamberlin Glacier shows that the 1958 climatic snowline (2,400 m) occurred where the mean July temperature was 4.0°C. At McCall Glacier, Benson, Fahl, Trabant, Well, and Wendler (1970) showed that the 1969 firn line was at an elevation of 2,210 m. Using their measured lapse rate of 0.57°C/100 m, the writer concludes that climatic snowline for that year was found where the mean July temperature was −1.5°C. Aerial photographs, topographic maps, and data provided by Péwé, Church, and Andreson (1969, table 1, p. 71–72) were used to determine that the mean July temperatures of 10 north-facing glaciers in the northern part of the east-central Alaska Range within 58 km of Delta Junction approximate 3.8°C. Thus in the Brooks Range and the northern flank of the east-central Alaska Range, the mean July temperatures at climatic snowline probably range at least from −1.5° to 6.8°C. This wide variation is undoubtedly a function of differing amounts of precipitation owing to differences in distance from precipitation source, wind drifting, and topographic effects. The writer thinks that the variation is too large and unpredictable to use for estimating the level of modern snowline in Alaska, unless no other data are available.

Methods Used In This Report

Several maps of modern and past snowlines have been compiled over the years by the writer and colleagues. Figure 11 is a map of Alaska showing distribution of isolines (isoglacihypses) of modern climatic snowline altitudes as determined from glaciers and perhaps a few perennial snowbanks. The points are located by plotting elevations of the lowest, small, north-facing cirque glacier from maps on a scale 1:63,360. Actually the points in each quadrangle are generally an average of the lowest cirques, thus eliminating an isolated single anomalous low point. Many of the central and western Brooks Range data are from maps on a scale of 1:250,000. Aerial photographs were used where necessary, especially for Grand Union Glacier on the Seward Peninsula. On the map, snowline was located at two-thirds of the elevation difference between the glacier toe and the upper ice limit on small circular cirque glaciers, or it was located where low medial moraines disappear beneath the snow cover on valley glaciers.

A few of the points plotted from the maps may be on perennial snowbanks and below modern snowline. Manley (1949, p. 185–186) concluded that in moist, cloudy, maritime areas such as Scotland, small snow beds may survive 450 m below firn line, as determined meteorologically. Ives (1960, p. 51) suggested that for continental climates the distance of snow patches below firn line could be between 763 and 825 m.

This method of snowline surface determination can be readily used in Alaska because good topographic maps are available for the entire State. However, the writer believes that although such a method gives a good representative pattern of modern snowline in Alaska, because of selecting the lowest cirque glaciers and perhaps also including a few snow patches, the snowline surface portrayed is perhaps slightly lower than it actually is.

Interpretation Of The Map

As Flint remarked (1971, p. 66), the snowline in three dimensions is the trace of a surface with complex undulations determined by temperature, precipitation, and forms of the land. The feature most clearly revealed by the map of modern snowline (fig. 11) is the characteristic rise from the windward coast of the State toward the interior—a broad effect of precipitation. This feature is also graphically demonstrated by the cross sections from west to east and south to north showing positions of snowline and the amount of precipitation (fig. 8, 9). In the central part of the State, the snowline rises steadily from a low of about 600 m in far western Alaska to about 2,000 m in adjacent Yukon Territory. Landward from the Gulf of Alaska, an abundant source of moisture, the snowline rises abruptly inward from about 400 to 600 m to the northwest, to the north, and to the northeast. It is interesting to see that moisture for the south side of the Alaska Range, even in the eastern part, apparently enters through the Cook Inlet trough.

Besides generally rising inland, the snowline rises over major topographic highs and forms a subdued replica of the mountain mass, owing to a decrease of precipitation with altitude, especially with the windward direction. If precipitation is insufficient for development of a complete icecap over a mountain mass, then small cirques and valley glaciers must form at various favorable sites up the slopes on all sides of the mountain range.

Perhaps the most interesting point illustrated by the map of modern snowline is the apparent lack of moisture for the glaciers contributed by the Arctic Ocean. Snowline rises from west to east along the axis of the Brooks Range and northern Alaska in general owing to

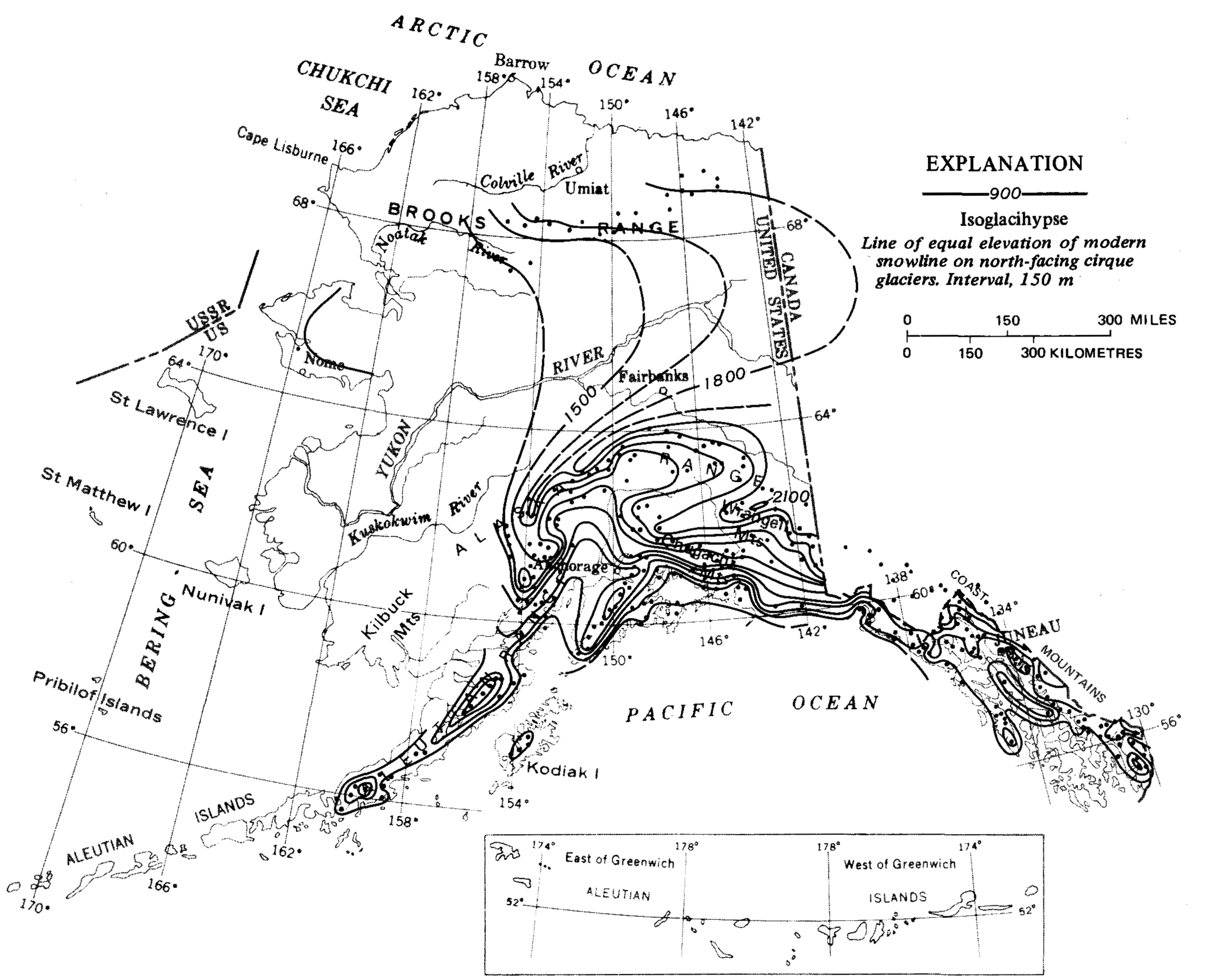


FIGURE 11.—Modern snowline surface in Alaska. Data points indicate the positions of north-facing glaciers found on maps at scales of 1:63,360 or 1:250,000. A total of 27 data points were mapped in the Yukon Territory, Canada, but 13 of these could not be shown as they lie within the area covered by the information block. Snowline was located at two-thirds of the elevation difference between the glacier toe and the upper ice limit on small circular cirque glaciers, or it was located where low medial moraines disappear beneath the snow cover on valley glaciers. The level of snowline on the small cirque glacier at the head of Grand Union Creek, western Seward Peninsula, was more precisely located using 1950 U.S. Navy aerial photographs. Compiled by R. D. Reger, 1970; data points in southeastern Alaska were plotted by R. H. Stinchfield, 1970.

moisture from the west. This relative lack of moisture contribution from the north today perhaps results from the perennial ice cover over most of the Arctic Ocean.

WISCONSINAN SNOWLINE

Methods of Determination

Position of snowline in Wisconsinan time can be estimated in many ways. One group of methods is concerned with utilizing the position of the cirque, another involves the reconstruction of the past equilibrium line by reconstructing past glacier surfaces, and another method develops a "glaciation limit" similar to the line determined with modern glaciers.

Wisconsinan snowline is commonly inferred to be the average altitude of floors of small cirques (Flint, 1957). Andrews (1965) placed the altitude at the junction of the cirque floor and the back wall. Another method for approximating former snowline is to follow the suggestion of Manley (1959) that snowline occurs at three-fifths of the altitudinal difference between the terminal and upper ice limit of a cirque glacier. Some place the Wisconsinan snowline at the median altitude between the terminal moraine of a given glacial advance and the highest point of a cirque headwall. Another method is to estimate snowline from the lower limit of icecaps. Richmond (1965) showed that for the Rocky Mountains, the late Pleistocene snowline depression obtained by using the figure for the median altitude between the terminal moraine and the highest point on the cirque wall is greater than using the figure obtained from cirque floors. He finds that the greatest snowline depression is recorded by a line approximated by the lower limit of icecaps.

It is possible to reconstruct a past "glaciation limit" using the same technique described for the modern snowline. Instead of mapping the summit altitudes of peaks containing small glaciers, one would map the summit altitude of peaks containing fresh cirques and the summit altitude of nearby mountains that have no cirques. This method would probably produce a surface that is actually above the past snowline surface but a generalized one quite acceptable for use. Flint (1971, p. 68) stated that using the "glaciation limit" produces a past snowline about 450 m higher than the actual past snowline.

Another technique for finding the position of past snowline is to determine the position of equilibrium line altitude on past glaciers using the accumulation area ratio (Porter, 1968, 1970), as mentioned under modern snowline. To apply this technique, the average accumulation area ratio of healthy glaciers (glaciers in dynamic equilibrium or advancing) in an area must be known, and the extent of the former ice limit must be accurately mapped so that the surface topography of the former glacier can be reconstructed. This technique applies only to symmetrical area-altitudinal distributions about the median altitude; piedmont glaciers cannot be used, nor can valley glaciers with large terminal bulbs.

Meier and Post (1962, p. 73) demonstrated that healthy glaciers have accumulation area ratios of 0.6 or more; therefore, as mentioned under the calculation for modern snowline, it is necessary to assume that the active past glaciers had an accumulation area ratio close to the 0.6 when calculating past equilibrium line altitude.

Determination of the elevation of snowline by this more accurate method puts the position of Wisconsinan snowline many metres lower than the position obtained for snowline using floors of empty cirques. In a study of Indian Mountain near Hughes, the writer determined the lowering of Wisconsinan snowline from modern snowline to be about 450 m on the basis of cirque floor studies (figs. 11, 12). R. D. Reger (in Reger and Péwé, unpub. data, 1974) calculated that Wisconsinan snowline was 585 m lower than modern snowline using the above equilibrium line altitude method. Therefore, snowline changes determined by differences in cirque floor must be considered minimum for valley glaciers.

Aside from the example from Indian Mountain, the writer knows of no application of this new method in Alaska. Calculation of the positions of Wisconsinan snowline in Alaska by this method must be made widely in the future and will be a strong step forward.

Methods Used In This Report

Because elevations of empty cirques can be easily plotted for most areas of Alaska and because all the information needed to calculate Wisconsinan snowline by Porter's method is not yet available, a surface representing the altitude of snowline in Alaska during Wisconsinan time (fig. 12) was compiled by plotting elevations of junction of the floor and back wall of sharp, empty, north-facing cirques from maps on a scale 1:63,360. Data for much of central and western Brooks Range are from maps on a scale of 1:250,000. Critical areas on the north side of the Brooks Range, Seward Peninsula, and on St. Lawrence Island were studied using aerial photographs. All maps used are modern compilations made by the U.S. Geological Survey using aerial photographs. The lowest sharp cirque was plotted in each 1:63,360 map that showed cirques except in southeastern Alaska where 593 cirques were plotted.

Inasmuch as it is difficult to differentiate between cirques of early and late Wisconsinan and Holocene ages without detailed studies and because some cirques were possibly reoccupied and deepened during two or more glaciations, the snowline surface is regarded as generalized Wisconsinan in age.

Another factor to be considered is tectonic displacement of cirques during and since Wisconsinan time. Many cirques along the edge of the Gulf of Alaska (especially Blying Sound, pl. 1) and along the south side of the Alaska Peninsula have been lowered or raised by tectonic movement. This movement accounts for the many cirques near or below sea level (figs. 13, 14) (Alpha, 1970).

Interpretation Of The Map

The surface of climatic snowline in Wisconsinan time in Alaska parallels modern climatic snowline everywhere at a lower elevation. The surface is lowest in the south and far west. There is an abrupt rise in the Gulf of Alaska from near sea level to about 1,500 m in interior Alaska and adjacent Yukon Territory. There is a steady, gradual rise from the Bering Sea on the west, eastward to Canada through central Alaska. In general, the depression of Wisconsinan snowline from modern snowline is about 300–400 m in the west and 450–600 m in the east (figs. 8, 9), but local variations occur. The vertical separation of the two snowline surfaces becomes greater toward the drier interior, as Hastenrath (1967, p. 546; 1971, p. 255) discovered in Peru.

The map supports the concept that the source of moisture in Wisconsinan time was the same as now, even

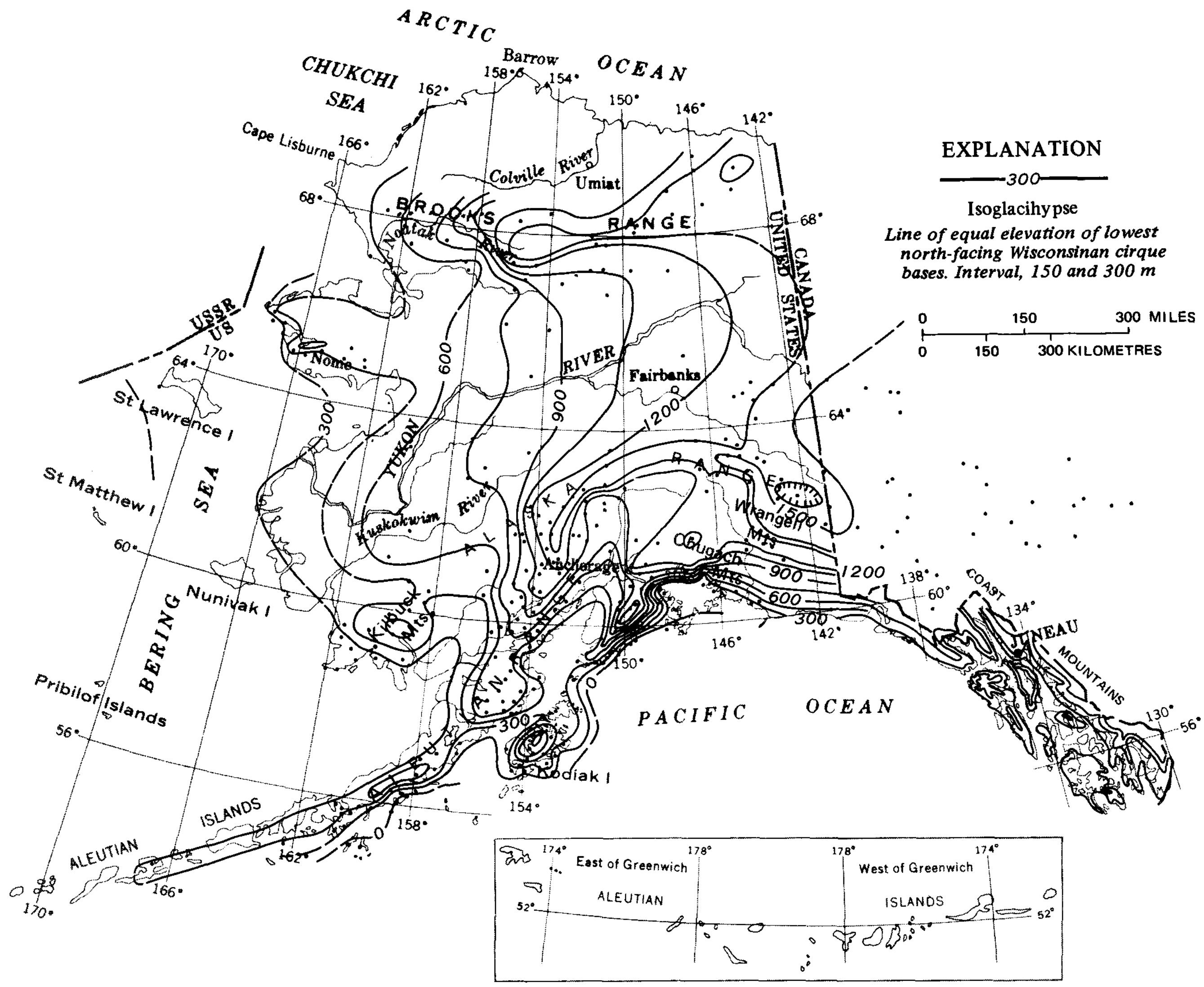


FIGURE 12.—Wisconsinan snowline surface in Alaska. Data points indicate the positions of the bases of the lowest, empty, north-facing cirques found on maps at scales of 1:63,360 or 1:250,000. Southward from Yakutat Bay only even-hundred-metre contours are shown owing to the steep gradients of the snowline surface. A total of 573 data points have been mapped in southeastern Alaska, but to preserve clarity, their locations are not shown. In adjacent Canada a total of 47 data points have been mapped, but not all are plotted; some lie within the area covered by the information block. Compiled by R. D. Reger, 1970. E. M. Schern plotted data points from southeastern Alaska, 1970.

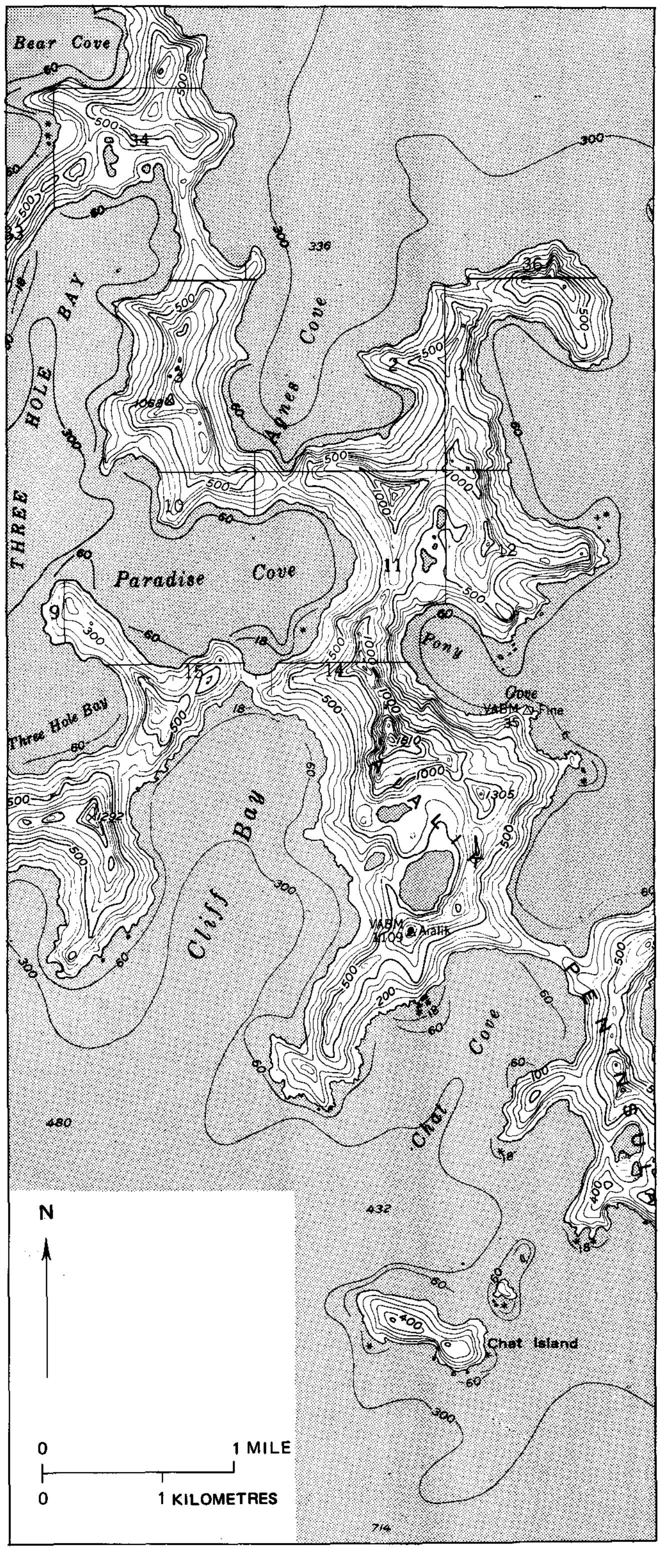


FIGURE 13.—Cirques drowned largely by tectonic lowering of the land. South side of the Kenai Peninsula. U.S. Geological Survey Blying Sound C–7, D–7 topographic quadrangles, 1950, 1952. Scale 1:63,360; contour interval on land 100 ft.

though the Bering Sea was smaller or perhaps absent during part of Wisconsinan time. The snowline distribution illustrates an apparent lack of significant moisture for glaciers contributed by the Arctic Ocean, which supports the suggestion that the Arctic Ocean has long been ice covered (Steuerwald and others, 1968) but does not support the concept of Ewing and Donn (1956, 1958; Donn and Ewing, 1966) that the Arctic Ocean was ice free and contributed abundant moisture to form ice sheets during Pleistocene time.

On the south side of the Chugach Mountains and the east side of the Kenai Mountains (figs. 13, 14) of southern Alaska, Wisconsinan snowline is detected as lying below present-day sea level (fig. 9). Such a lowering of cirques is the result of tectonic activitiy since Wisconsinan time (Plafker and Rubin, 1967; Plafker, 1969, p. I58). In 1964 parts of the area were further warped downward by the March 27 earthquake (Plafker, 1965). The snowline study indicates that except in certain tectonically active areas, Wisconsinan topography, wind belts, and precipitation were generally similar to what they are today. It is perhaps now timely to consider a position reconstruction of the climatic pattern and atmospheric circulation in Alaska during Wisconsinan time by statewide information on snowline depression.

HOLOCENE GLACIATIONS

Events that took place less than 10,000 years ago are assigned in this report to the Holocene Epoch. The date 10,000 B.P. (before present) has been adopted as the beginning of the Holocene by the International Association for Quaternary Research (INQUA). At about 10,000 B.P., a worldwide warming occurred, recorded in marine deposits, the pollen spectrum, and snowline position (Fairbridge, 1968, p. 528–530; Morrison, 1969). A warming is recorded at this time in northwest Alaska (McCulloch and Hopkins, 1966) and in central Alaska (Péwé, 1975). Two or possibly three glacial cycles are recorded in many parts of Alaska during this interval. The early cycle took place between 8,500 and 6,000 years ago in some areas. A later series of glacial cycles began about 4,000 years ago and ended only a century ago, but the multiple early and late glacial advances within this interval may have been separated by a significant deglacial interval of unknown duration approximately 1,500–2,000 years ago.

The Holocene glacial record as recognized throughout much of central and northern Alaska is typified by the sequence recorded by Porter (1964) in the Anaktuvuk Pass area of the Brooks Range. Glaciers were there in an advanced position from about 8,300 years ago until about 6,300 years ago (table 2), during Porter's Anivik Lake Stade of the Itkillik Glaciation (as interpreted by Porter). A long deglacial interval was followed by the

Alapah Mountain Glaciation (as interpreted by Porter), which culminated about 2,800 years ago. Finally, the double moraines of the Fan Mountain Glaciation, which generally lie near or in the cirques or near small remnant glaciers, are believed to have formed during the last several centuries.

Karlstrom (1964, pl. 7, p. 44, 56, 64) reported a somewhat similar sequence on the Kenai Peninsula of southern Alaska. The last readvance during the Tanya Stade of the Naptowne Glaciation is thought by Karlstrom to have taken place about 6,000 years ago. Logs buried in drift indicate that some glacial advances during the

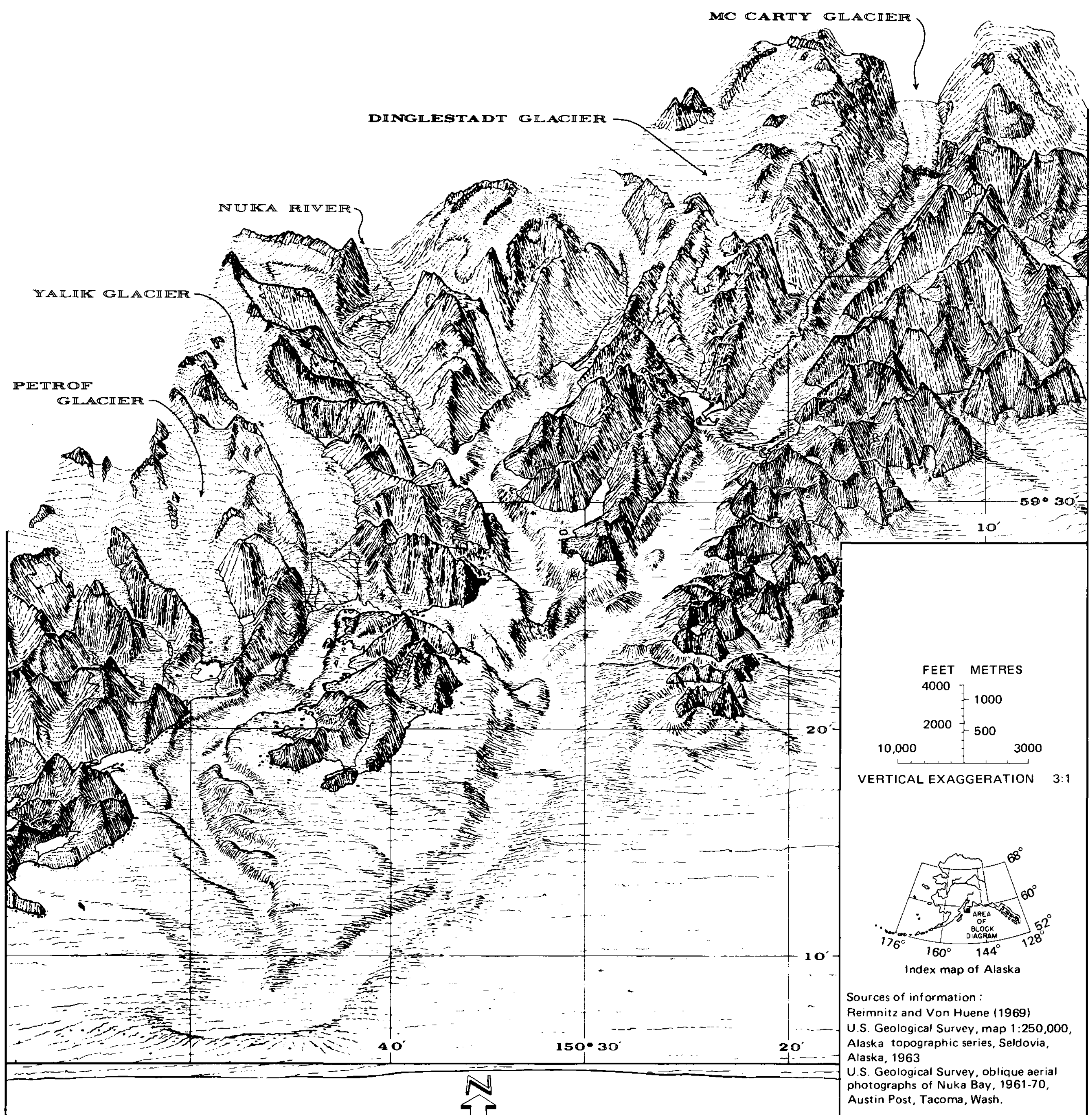


FIGURE 14—Orthographic drawing of Nuka Bay, Alaska, south side of Kenai Peninsula, showing drowned cirques and glacial valleys. Drawing by Tau Rho Alpha, U.S. Geological Survey, 1970.

Tustumena Stade of the Alaskan Glaciation took place more than 2,500 years ago, and indirect evidence led Karlstrom to conclude that the Tustumena Stade extended from about 5,000 to 2,000 years ago. The Tunnel Stade, also of the Alaskan Glaciation and well dated by radiocarbon analyses of logs buried in the older moraines and by ring counts from trees growing on young moraines, consists of several glacial advances during the last 1,500 years.

Records of early Holocene glaciations along the coast of the Gulf of Alaska and in southeastern Alaska are lacking. Plafker and Miller's careful study (1958) of the moraines of the Malaspina and neighboring glaciers records a glacial advance that culminated between 1,400 and 700 years ago and that was more extensive than any Wisconsinan or early Holocene glacial advance. A glacial recession followed, succeeded by a readvance that began before A.D. 1700 and that ended in different places between A.D. 1791 and A.D. 1904. Studies of various parts of southeastern Alaska also suggest that glaciers advanced farther during the last few centuries than at any previous time within the last 10,000 years (Lawrence, 1950, 1958; Heusser and Marcus, 1964). The most detailed record is provided by Goldthwait's study of Glacier Bay (1963, 1966). A prolonged process of filling the bay with outwash gravel began about 7,000 years ago. Glaciers were present in the mountains but were lacking in the fiords and lowland valleys of the Glacier Bay area; glaciers in southeastern Alaska then withdrew farther than where they are today. Forest beds and beds of sand apparently record a slackening of sedimentation and perhaps an accelerated or renewed glacial retreat during a few centuries after 2000 B.C., but by 1500 B.C., glaciers had begun to advance again, and during the ensuing three millennia they filled the entire bay. Finally, shortly before A.D. 1791, a rapid glacial recession began that has continued essentially without interruption to the present day. This catastrophic retreat has resulted in recessions of the major ice tongues through distances of 75–100 km.

Work by Williams and Ferrians (1961) in the upper Matanuska River valley reveals a history of Holocene advances of the Matanuska Glacier supported by radiocarbon dating (table 2). After its retreat in late Wisconsinan time, the glacier readvanced 4–8 km more than 8,000 years ago and then receded. After local canyon cutting by streams, the glacier readvanced 1.6 km a few thousand years ago. Moraines a few hundreds of years old lie near the glacier front.

Detailed work using dendrochronology and lichenometry enabled Péwé and Reger (Péwé, 1951a, 1957b; Reger, 1964, 1968; Reger and Péwé, 1969) to date recent advances of five glaciers in the central part of the Alaska Range. Advances were dated as A.D. 1830, 1875, 1650(?), 1580(?) and pre-1580(?). An inspection of aerial photographs shows that fresh moraines similar to those dated as A.D. 1830 and 1650(?) are widespread throughout Alaska. An unpublished report by James Rowe of Arizona State University on dendrochronologic and lichenometric dating of Holocene moraines of Puget Glacier on the southeast side of the Kenai Peninsula indicates two prominent moraines that probably date from 1830 and 1650(?).

Recent summaries of the glacial history of the last 5,000 years in the North American Cordillera are by Porter and Denton (1967) and Viereck (1968).

SUMMARY

Alaska contains one of the longest and best records of late Cenozoic glaciations in the world, extending from late Miocene time until today and including both terrestrial and marine glacial sediments. The deposits in some areas may be tied into known marine deposits and correlated worldwide. Early and middle Pleistocene deposits not associated with the sea are correlated with difficulty, but Illinoian, Wisconsinan and Holocene deposits are widespread and dateable.

Glaciers in the past occupied about 50 percent of Alaska and today cover only 5 percent. During glacial maximums, snowline was lowered many hundreds of metres, with Illinoian snowline about 170 m below Wisconsinan snowline. The Wisconsinan snowline parallels the modern snowline everywhere at a lower elevation. A steady rise of modern and Wisconsinan snowlines from the west and south to the east and northeast reflects the broad effect of precipitation on snowline. There is no evidence for nourishment of glaciers by moisture from the Arctic Ocean.

Glaciation in Wisconsinan time was complex, and deposits of at least two major glacial advances are widespread. Glacial advances of the last 10,000 years are excellently displayed in Alaska, and dendrochronology and lichenometry will be useful for differentiating widespread recent advances from erratic glacial surges.

EOLIAN DEPOSITS

Eolian deposits cover a large part of the low-lying areas of Alaska and are mostly of middle to late Pleistocene age; only a few deposits are Holocene. The deposits include large areas of loess and reworked loess, smaller areas of stabilized dunes, and very small areas of active dunes. Practically all loess was derived from glacial outwash plains, and the few areas where loess is accumulating in significant quantities today adjoin the braided flood plains of silt-laden glacial streams. Most stabilized sand dunes also adjoin sandy, gravelly glacial outwash plains of late Pleistocene age, but a large area of stabilized dunes on the Arctic Coastal Plain and a small area of stabilized dunes on Adak Island in the Aleutians (Judson, 1946) probably accumulated on the lee side of extensive sandy beaches. Most of the presently active dune areas in Alaska adjoin low-lying sandy beaches, but two large active dune areas lie inland within larger areas of stabilized dunes.

SAND DUNES

Many of the older reports on the regional geology of areas in Alaska note the presence of active dunes, but the much larger areas of stabilized dunes commonly escaped notice. Not until 1951 was the first small-scale map published showing the distribution of active and stabilized sand dunes (Black, 1951a), and four additional large areas of dunes have been recognized within the ensuing 10 years (fig 15). The areas of stabilized dunes are not easily recognized because they commonly are heavily vegetated, and until aerial photographs came into wide use, their distinctive form was not obvious.

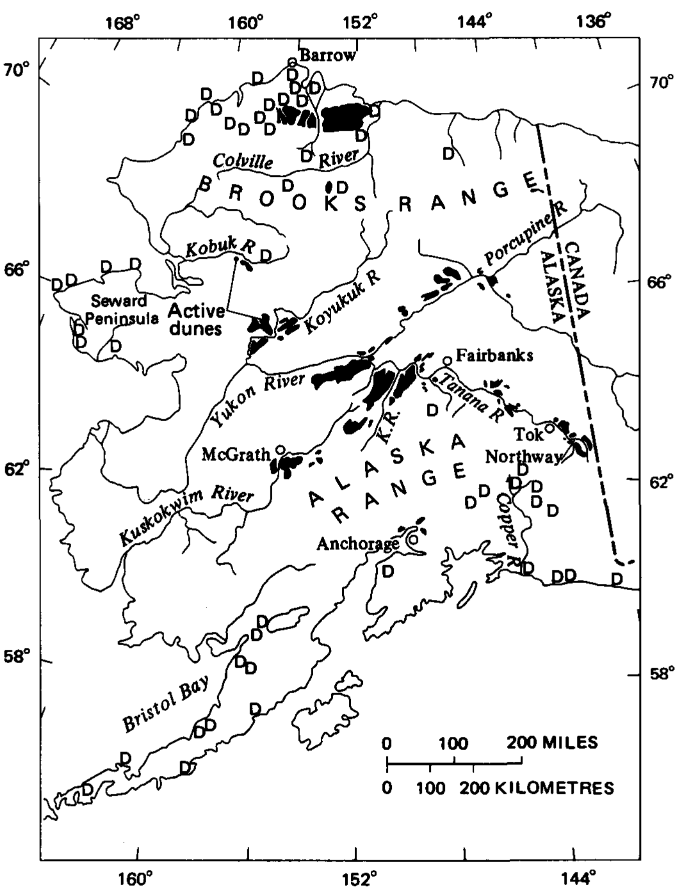


FIGURE 15.—Major areas of stabilized sand dunes in Alaska. Most coastal dunes as well as small parts of the stabilized sand dune areas on the Koyukuk and Kobuk Rivers are active today. Symbols: D, small areas of sand dunes; K. R., Kantishna River. Compiled by F. R. Weber from published and unpublished sources, 1961.

ACTIVE DUNES

Active dunes on the coastal beaches are 0.3–18 m high and consist mainly of blowout and transverse dunes, foredunes, and cliff-head dunes. They are known mainly from the Yakutat area (Miller, 1961a), the delta of the Copper River (Tarr and Martin, 1914, p. 466), the Bristol Bay area (Black, 1951a, p.103), Seward Peninsula (D. M. Hopkins, unpub. data, 1957), and northern Alaska (Black, 1951a) but are present everywhere on sandy coasts. The two largest areas of active dunes in Alaska are not, however, along modern beaches but are in the Koyukuk and Kobuk River valleys. The active Nogahabara dune field in the lower valley of the Koyukuk is only a small part of a much larger Pleistocene dune field that is now mostly inactive. The active dune area is approximately 65 km² and is composed mainly of transverse dunes 15–60 m high and 90 m long or more. The middle Kobuk valley contains another large area of Pleistocene dunes, of which 130 km² contains active dunes (Fernald and Nichols, 1953; Fernald, 1964).

STABILIZED DUNES

Stabilized dunes, probably of Illinoian and Wisconsinan age, are widespread, especially in central and northern Alaska, but only locally have they been mapped in even a reconnaissance fashion. Most of the dunes occur in the valleys of the Tanana, Yukon, Kuskokwim, Koyukuk, and Kobuk Rivers—rivers that were important streams draining glaciated areas in the past.

ILLINOIAN

An area of about 780 km² of poor to fairly well formed sand dunes lies against the bedrock hills up to an elevation of 520 m north of the Tanana River in the vicinity of Big Delta, about 145 km southeast of Fairbanks. These dunes are parabolic and have rather subdued to fairly fresh relief. Local relief is from 0.3 to 3 m. The dunes are thought to have originated in Illinoian time (table 3) when winds blew sand northward from the widespread outwash plains and glacial flood plains of the Delta and Tanana Rivers (Péwé, 1965b, p. 46 –51). Most of the sand forms a blanket covering the hills and has no dune form. The sand blanket and dunes subsequently have been dissected and covered with loess 0.3–2 m thick (fig. 16).

The largest area of stabilized dunes in Alaska extends from the Tanana River southward for 160 km into the drainage of the upper Kuskokwim River (fig.1). These dunes, which are thought to be Illinoian in age, occur on high terraces of the Tanana River and its tributaries from the south, especially the Kantishna River. Several masses of sand dunes, which total 15,500 km² (fig.16), were mapped by F. R. Collins (unpub. data, 1958) and Péwé, Wahrhaftig, and Weber (1966). The dunes are

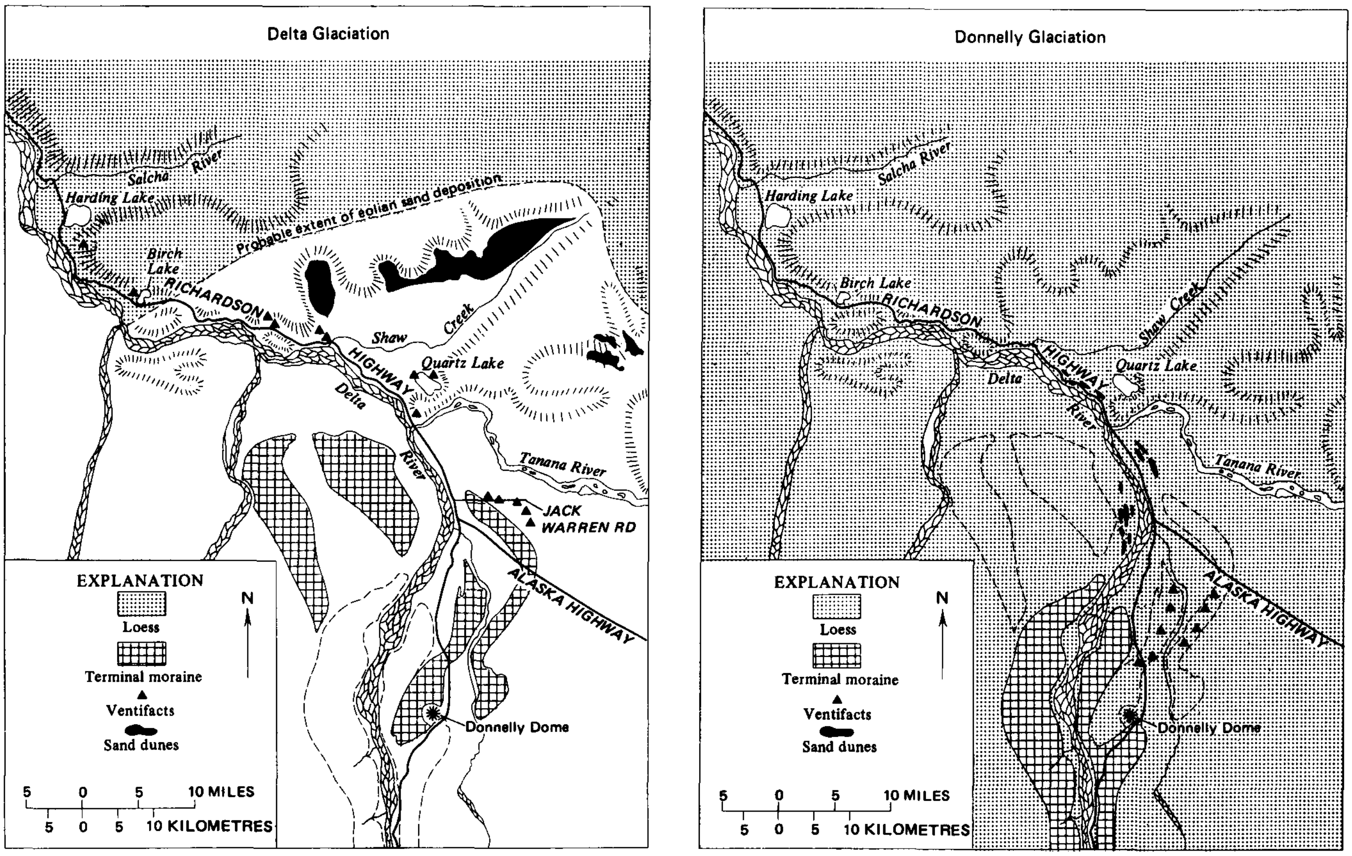


FIGURE 16.—Distribution of loess, sand dunes, and ventifacts in the Big Delta area, Alaska, in the time of the Delta Glaciation (Illinoian) and Donnelly Glaciation (Wisconsinan) in relation to the glacial advance. From Péwé (1965b).

well formed and have a local relief of 2–60 m. In most of the areas, they occupy the lowland at an elevation of 100 m; elsewhere, they lap up against the bedrock hills and are well formed up to 300 m above sea level. These stabilized dunes are dominantly longitudinal but locally have a parabolic form. Like others in interior Alaska, they are covered with spruce, birch, and aspen. Dune sand was blown by northeast winds from the glacial outwash plains of such rivers as the Tanana, Kantishna, Toklat, and others draining the Alaska Range. The dunes are covered with a few centimetres to a few metres of loess.

An area of 500 km² of gently rolling sand ridges occurs in the valley of the middle Kuskokwim River (Fernald, 1960). These dunes were derived from sediments of outwash plains and are related to the Selatna Glaciation (tables 2, 3). Fernald believed they are Wisconsinan and pre-Wisconsinan in age; however, the writer believes that they are Illinoian in age because of their association with the Selatna Glaciation.

Vegetated, topographically rather subdued dunes, largely of the parabolic type, occupy an area of 3,200 km² on high terraces in the lower Koyukuk River valley (Weber and Péwé, 1961). These dunes, formed by northeast winds blowing across outwash deposits, are probably pre-Wisconsinan in age (table 3).

Dunes that are now stabilized were deposited by wind blowing across glacial outwash in the middle Kobuk River valley (Fernald, 1964). These dunes cover approximately 600 km² of high river terraces and may have originated as early as late Illinoian time but no doubt were reactivated and enlarged during Wisconsinan time.

WISCONSINAN AND POST-WISCONSINAN

Well-formed parabolic and compound dunes of Wisconsinan age (table 3) cover approximately 1,150 km² on high terrace remnants of the upper Tanana valley (Fernald, 1965b). These dunes are 30–60 m high and are best developed in the vicinity of Northway and Tok (fig. 15). Much of the sand blanket laps up on the bedrock hills to the north of the present river valley. The sand is gray to black and is derived from the wide glacial valley trains and outwash plains. The dune areas are much dissected and covered with a thin blanket of loess.

Fernald (1960) described an area of about 800 km² of well-formed parabolic dunes of Wisconsinan age on high terraces in the middle Kuskokwim River valley. About 600 km² of eolian sand of Wisconsinan or greater age covers most of the upland marginal to the Yukon Flats (Williams, 1962), and eolian sand of Wisconsinan age covers approximately 500 km² of the Yukon Flats.

Dunes are widespread in an area of 13,000 km² on the Arctic Coastal Plain west of the Colville River (fig. 15). In this region large areas of stabilized longitudinal, parabolic, and multicycle dunes and local areas of active dunes occur along most rivers and streams (Black, 1951a, p. 92–96; Walker, 1967). Stabilized dunes generally have a longitudinal and parabolic form. Most of the longitudinal dunes are less than 1 km long, but the longest is about 2.4 km. These sand deposits are as much as 3–6 m thick. Black believes that these stabilized dunes were formed in Holocene time by northeast winds, the same direction as the present prevailing winds that formed the active dunes along the streams (Rickert and Tedrow, 1967). There have been many incidental observations on the sand dunes of northern Alaska during the geomorphic studies of the past 10 years (Rosenfeld and Hussey, 1958; Livingstone, 1954).

Stabilized dunes cover a belt several miles wide on the east and northeast sides of St. Paul Island. The dunes are of the transverse type but have been slightly modified by secondary development of small paraboloid dunes prior to complete stabilization. Dune crests stand 3–5 m above interdune areas. The dunes were formed by winds blowing from the east and northeast. The dunes cover a Pelukian (Sangamon) (table 3) shoreline and lie on pillow lavas of Sangamon age; they contain ice wedge casts, and a thick soil profile is developed on them. The sand contains abundant quartz and mica minerals that are not found in the basaltic lavas of St. Paul Island. These observations indicate that the dunes formed during the Wisconsinan Glaciation at a time when sea level was low, the Bering shelf exposed, and a large braided river—probably the Yukon—was delivering large quantities of sediment to a shoreline nearby (D. M. Hopkins, written commun., Feb. 3, 1968).

VENTIFACTS

Ventifacts, stones abraded by wind-blown sand were reported in Alaska (Smith and Mertie, 1930, p. 249; Black, 1951a, p. 108–109; Péwé, 1965b) but are not widespread. Well-developed ventifacts were reported by D. M. Hopkins (written commun., 1968) on Black Diamond Hill on St. Paul Island in the Pribilofs. These ventifacts, which are as much as 2 m in diameter, are Holocene and are still being formed today. The most widespread and best developed ventifacts known in Alaska occur in the glaciated Big Delta area and in the adjacent unglaciated terrain in the middle Tanana valley. Here the ventifacts of both Illinoian and Wisconsinan age have been useful in interpreting the geologic history of the area (Péwé and others, 1953; Péwé and Holmes, 1964; Péwé, 1965b, p. 47–52) (fig. 16).

LOESS

HISTORY

Loess probably is the most widely distributed sediment of Quaternary age in Alaska. It forms a blanket, ranging in thickness from a few millimetres to more than 60 m, that covers almost all areas that lie below altitudes of 300–450 m. Thick deposits of loess are most widely distributed in central and western Alaska (fig. 17) (Péwé, 1968a). Most of the loess was deposited during Illinoian and Wisconsinan time, but wind-blown silt is still being deposited in many areas (Péwé, 1951a). The loess is of great economic importance because it forms the soils of the principal agricultural areas in Alaska (Bennett and Rice, 1919; Péwé, 1954, 1958b; Trainer, 1961).

The wind-blown silt of Alaska was first described by Spurr (1898 p. 200–211), who applied the name "Yukon silts" to deposits in the valleys of the Yukon River and its tributaries. For nearly 50 years, the origin of these silts was a matter of controversy; various workers argued for a fluvial origin, a lacustrine origin (Spurr, 1898, p. 200–230; Prindle, 1913, p. 50; Eakin, 1916, p. 73, 1918, p. 45; Dorsh, 1934, p. 4), a marine and estuarine origin (Harrington, 1918, p. 30), or a residual origin (Wilkerson, 1932, p. 18–19). Although Tuck (1938) and Eardley (1938) correctly recognized that most of the silt was of wind-blown origin and Capps (1940) referred to eolian origin, Taber (1943, p. 1471–1479; 1953, p. 321–336; 1958) argued that the silt, including the reworked valley-bottom silt rich in organic material, was entirely the product of prolonged disintegration of local bedrock by frost riving. The writer, however, (Péwé, 1955) showed that the thick silt deposits of the Fairbanks district are of eolian origin, and Trainer (1961) and Hopkins (1963) confirmed the eolian origin of surficial silt blankets in the Matanuska Valley and central Seward Peninsula.

ORIGIN AND PHYSICAL CHARACTERISTICS

The loess of Alaska was blown from the vegetation-free flood plains of braided glacial rivers; consequently, loess is thickest near streams draining glaciated areas. At the present time, loess is being deposited most rapidly near the modern outwash streams. Loess has been deposited on ridges as high as 760 m above sea level, but most seems to have been deposited at altitudes of less than 450 m. A large part of the loess falling on summits and slopes of hills has been washed into valley bottoms to form thick deposits of bedded to massive silt that is rich in organic debris. These deposits locally are called muck (Péwé, 1952a). The bedded character of the valley-bottom silt has been used as evidence to support hypotheses for the marine, lacustrine, and residual origin of loess.

The loess in Alaska is similar in texture and color to loess elsewhere in the world. Typically, 80–90 percent of the particles are between 0.5 and 0.005 mm in size (fig. 18). Particle size increases with decreasing distance from the river flood plains that constitute the source areas (Trainer, 1961; Péwé and Holmes, 1964; Davidson and others, 1959; Lindholm and others, 1959, p. 49). Colors range from chocolate brown through tan to olive gray and neutral gray. Mineral and chemical composition reflect differences in bedrock composition of the glaciated valleys from which the particles were ultimately derived. Loess typically contains minerals that are not present in the bedrock of areas where it was deposited by the wind (Péwé, 1955; Hopkins, 1963).

Textural studies by Dement (1962) indicated no im-

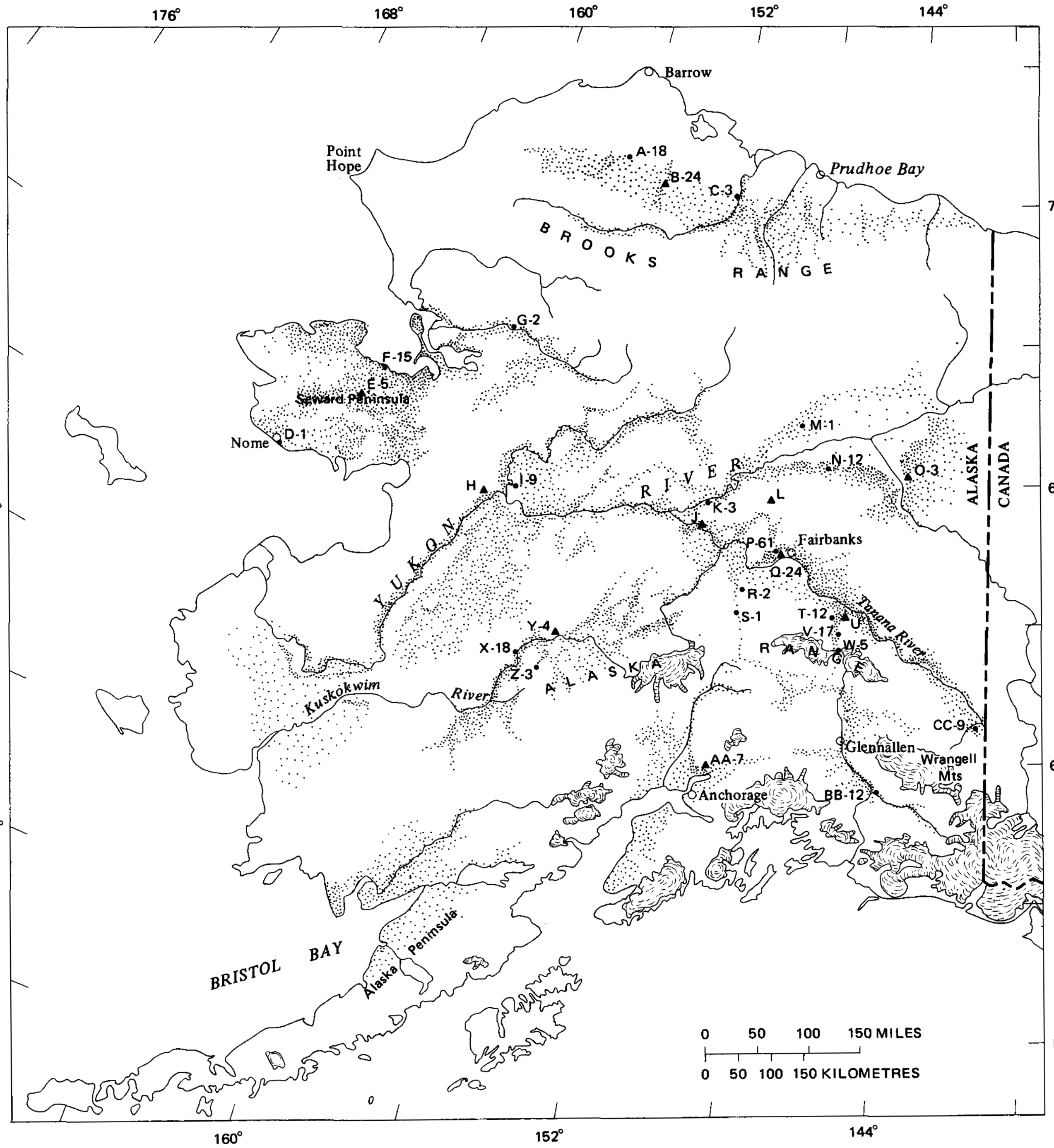


FIGURE 17.—Loess deposits of Alaska. General thickness indicated by density of stippling. Compiled by T. L. Péwé, 1961.

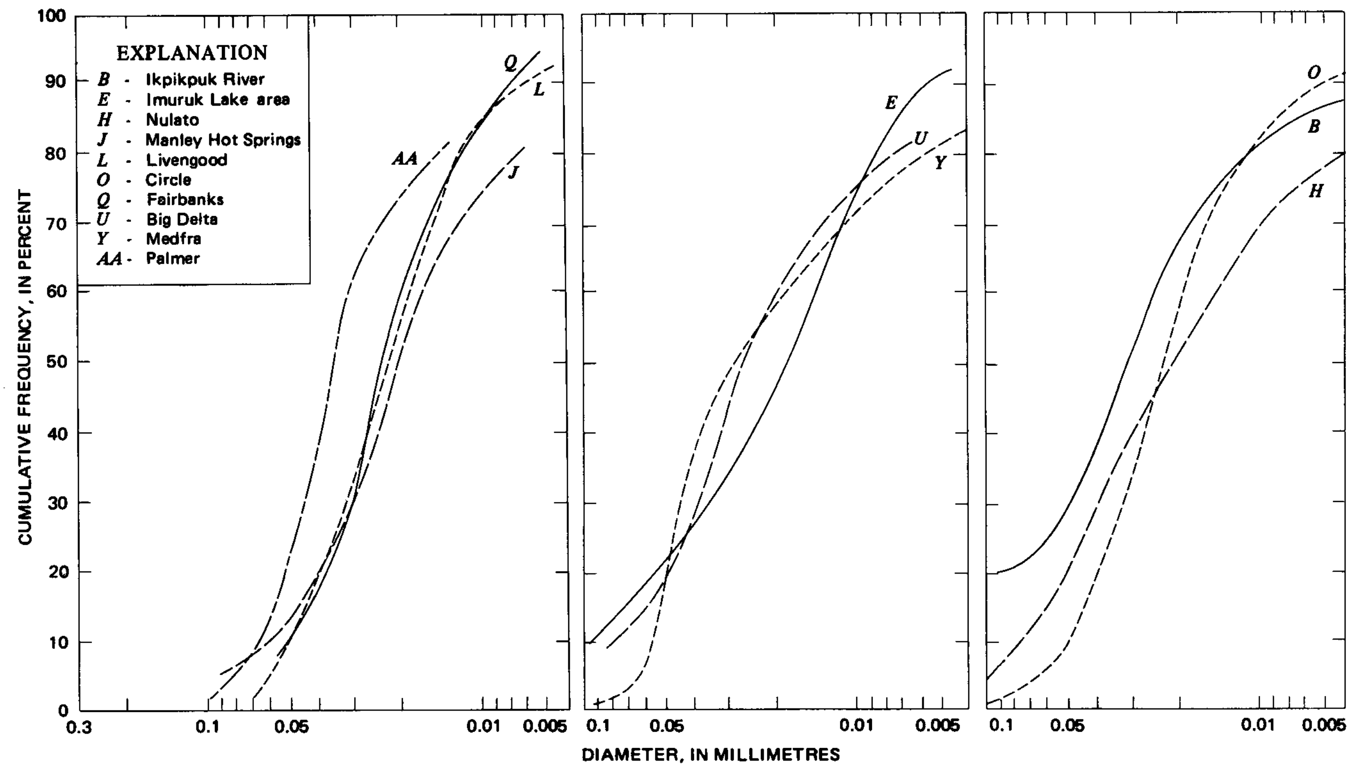


FIGURE 18.—Cumulative-frequency grain size curves of loess from various locations in Alaska. Letters refer to locations in figure 17 (map).

portant changes in clay content throughout the Fairbanks Loess. Thin textural bands high in fresh clay-sized particles occur parallel to the surface. The textural bands are not generic and not part of a soil development. They probably are depositional, the result of slight changes in source and deposition of loess. Frost action also has been suggested as a process involved in the concentration of clay.

Source of data for specific localities cited

[Symbols:▲3, thickness in metres; mechanical analyses shown in figure 18.●15, thickness in metres]

A–● R. F. Black (written commun., 1961).
B–▲ P. V. Sellmann (oral commun., 1961).
C–● R. F. Black (written commun., 1961).
D–● D. M. Hopkins (unpub. data, 1961).
E–▲ Do.
F–● Do.
G–● A. T. Fernald (written commun., 1961).
H–▲ T. L. Péwé (unpub. data, 1961).
I–● Do.
J–▲ T. L Péwé (1955, p. 708).
K–● J. R. Williams (written commun., 1961).
L–▲ T. L. Péwé (1955, p. 707).
M–● J. R. Williams (written commun., 1961).
N–● Do.
O–▲ Do.
P–● T. L. Péwé (unpub. data, 1961).
Q–▲ T. L. Péwé (1955, p. 704).
R–● Clyde Wahrhaftig and R. F. Black (1958, p. 87).
S–● Do.
T–● T. L. Péwé and G. W. Holmes (1964).
U–▲ T. L. Péwé (1955, p. 707).
V–● T. L. Péwé and G. W. Holmes (1964).
W–● T. L. Péwé (unpub. data, 1961).
X–● A. T. Fernald (1960, p. 247).
Y–▲ A. T. Fernald (1960, p. 247-249).
Z–● A. T. Fernald (1960, p. 247).
AA–▲ F. W. Trainer (1961, p. 14).
BB–● D. R. Nichols (written commun., 1961).
CC–● A. T. Fernald (written commun., 1961).

FIGURE 17.—Continued.

The carbonate content differs from place to place, both as a result of differences in the amount of carbonate rocks in the ultimate source areas and as a result of local differences in the history of postdepositional weathering. Near Fairbanks the freshest loess has no appreciable quantities of carbonate, but unweathered loess near Manley Hot Springs, 130 km to the west, effervesces vigorously upon application of hydrochloric acid. Most loess is highly micaceous, but Fernald (1960, p. 250) found very little mica in loess in the upper Kuskokwim River valley. The few mineral and chemical analyses of Alaskan loess that have been published are from the Fairbanks area (Péwé, 1955), Palmer and Big Delta (Stump and others, 1959, p. 17; Lindholm and others, 1959, p. 53), the upper Kuskokwim River valley (Fernald, 1960, p. 250), and the Imuruk Lake area of the Seward Peninsula (Hopkins, 1963).

As is typical of loess throughout the world, Alaskan loess is massive, cliff forming, and locally perforated with swallow-nesting tunnels. Only in some areas of Alaska does the loess contain fossils of air-breathing mollusks. In the great loess areas of the interior of Alaska, especially the Tanana and Yukon valleys, the loess, though examined in hundreds of exposures, has yielded fossils of air-breathing snails in only the very young loess near Big Delta (Péwé, 1955) and in loess of probable Wisconsinan age near Tofty (Repenning and others, 1964). However, the loess near Chitina and Gulkana in the Copper River Basin contains many pulmonate snails, as does the loess near Fort Hamlin in the middle Yukon River valley (Williams, 1962). In the Kotzebue Sound area, air-breathing and aquatic snails and clams were collected from the deposits of thaw lakes that existed during loess deposition (McCulloch and others, 1965).

DISTRIBUTION AND THICKNESS

The extent of loess in Alaska and thickness data at selected localities are presented in figure 17. The thickest loess known in Alaska is north of the Tanana River near Fairbanks, where a blanket 61 m thick covers the top of Gold Hill (Péwé, 1955). Greater thicknesses of silty sediments (as much as 95 m) occur in the bottoms and lower slopes of small valleys in the Fairbanks area, but these represent deposits reworked from adjoining slopes. Deposits 3–12 m thick occur along the entire north side of the Tanana valley and on both sides and in the middle of the lower Yukon River valley (Eardley, 1938). Loess is thick adjacent to the Yukon Flats (Williams, 1955, p. 125; 1962) and the upper Kuskokwim valley (Fernald, 1960, p. 247) (fig. 17). Many metres of wind-blown silt exist along the central Copper and lower Chitina valleys and near Palmer (Stump and others, 1959). Loess more than 1.5 m thick covers about three-quarters of the area of Seward Peninsula (Hopkins, 1963, pl. 3). The major river valleys on the Arctic Slope are flanked by loess that is generally many metres thick, but the loess blanket thins considerably immediately away from the rivers (Smith and Mertie, 1930, p. 249; Black, 1951a, p. 92; Detterman and others, 1963, p. 304–305). Loess on the Pribilof Islands (outside area mapped in fig. 17) averages 15 cm thick, and some slopes on St. George Island have 2 m of loess. This loess probably originated as silt blown from the extended flood plains of the Yukon and Kuskokwim Rivers across the Bering platform in Illinoian and Wisconsinan times (D. M. Hopkins, unpub. data, 1965; Scholl and others 1968, p. 320). Deposits a few centimetres to a few metres thick are known locally elsewhere in Alaska (Black, 1951a; Muller, 1955, p. 132; Karlstrom, 1955, p. 133; Hoare and Coonrad, 1959; Coonrad, 1957; Hoare and Condon, 1971; Chapman and others, 1971) (fig. 17).

It is interesting to note that loess is said to be absent, or nearly so, from St. Lawrence Island. J. M. Hoare, W. H. Condon, D. S. McCulloch, and D. M. Hopkins (written commun., 1968) confirmed this, for example, in the vicinity of Savoonga on the north side of the island. These observations perhaps suggest that the position of the major rivers across the Bering shelf in glacial times may not have been close to the island. This suggestion should be added to the growing body of evidence shedding light on the environment and glaciation of the Bering shelf, a subject of current controversy and investigation.

A thin layer (about 10 cm) of loess of Holocene age seems to be present near Barrow, but loess of Wisconsinan time apparently is lacking. Brown (1965a) believed that the Holocene silt is loess and was derived from drained lake basins; however, Carson (1968, p. 25) stated the silt does not represent "wind-laid deposit, but simple reworked Gubik sediments winnowed from lighter weight organic fines by normal wave processes."

Detailed studies were made in some localities regarding the thicknesses of loess. Hundreds of measurements were taken in the Fairbanks area (Péwé, 1955, 1958b; Péwé and Rivard, 1961; Williams and others, 1959), and isopach maps were made for the Matanuska Valley (Trainer, 1961) and for the lower reaches of the Delta River valley (Davidson and others, 1959; Péwé and Holmes, 1964). In all instances, the loess is thickest near the rivers.

STRATIGRAPHY

The periods of greatest loess deposition in Alaska correspond to glacial maximums. During these periods the source areas, vegetation-free flood plains and outwash plains, were at their greatest extent. In the periglacial areas, such as the Tanana, Yukon, and Kuskokwim valleys, repeated glacial advances with related wind conditions and source areas favorable for loess deposition have created multiple loess blankets (table 3) (Péwé, 1955; Fernald, 1960; Péwé, 1965b). At Cape Deceit near Deering (pl. 1) on the south shore of Kotzebue Sound, Guthrie and Matthews (1971) recorded thin loess layers of pre-Illinoian, Illinoian, Wisconsinan, and Holocene age. This is the oldest loess reported in Alaska.

Detailed stratigraphic studies of loess are limited to the Seward Peninsula and adjoining area (Hopkins, 1963; McCulloch and others, 1965; Guthrie and Matthews, 1971) (fig. 19) and to the central Yukon-Tanana Upland (Péwé, 1952a, 1955, 1958b).

One of the most interesting records of Quaternary events in Alaska comes from an interpretation of the loess deposits and underlying creek gravel near Fairbanks (table 3). Extensive exposures created by placer mining operations have revealed a complicated history of loess deposition alternating with periods of fluvial erosion.

In Illinoian time, the hills were blanketed with loess derived from the flood plain of the Tanana River and glacial outwash plains south of the Fairbanks area. Much of this windblown silt was retransported to creek valley bottoms, incorporated much organic debris, in-

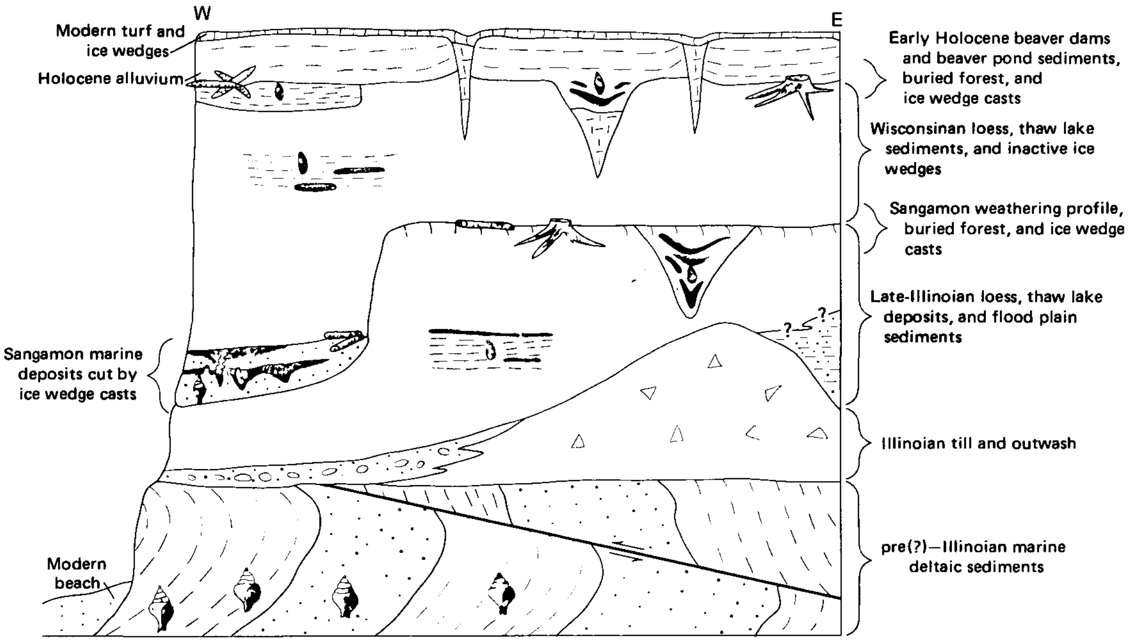


FIGURE 19.—Diagrammatic cross section of Baldwin Peninsula, Alaska, showing Quaternary deposits. From McCulloch (1967).

cluding vertebrate remains, and became perennially frozen. Indications of the antiquity of this loess are (1) its position unconformably beneath a younger silt deposit, the base of which is older than 56,900 years (Matthews, 1968b, 1970) (fig. 20), (2) joints that were heavily stained by iron oxide and cemented before the deposit became perennially frozen, and (3) evidence from fossil ice wedges that the loess, once frozen, was thawed, then again perennially frozen.

This period of loess deposition was followed by an

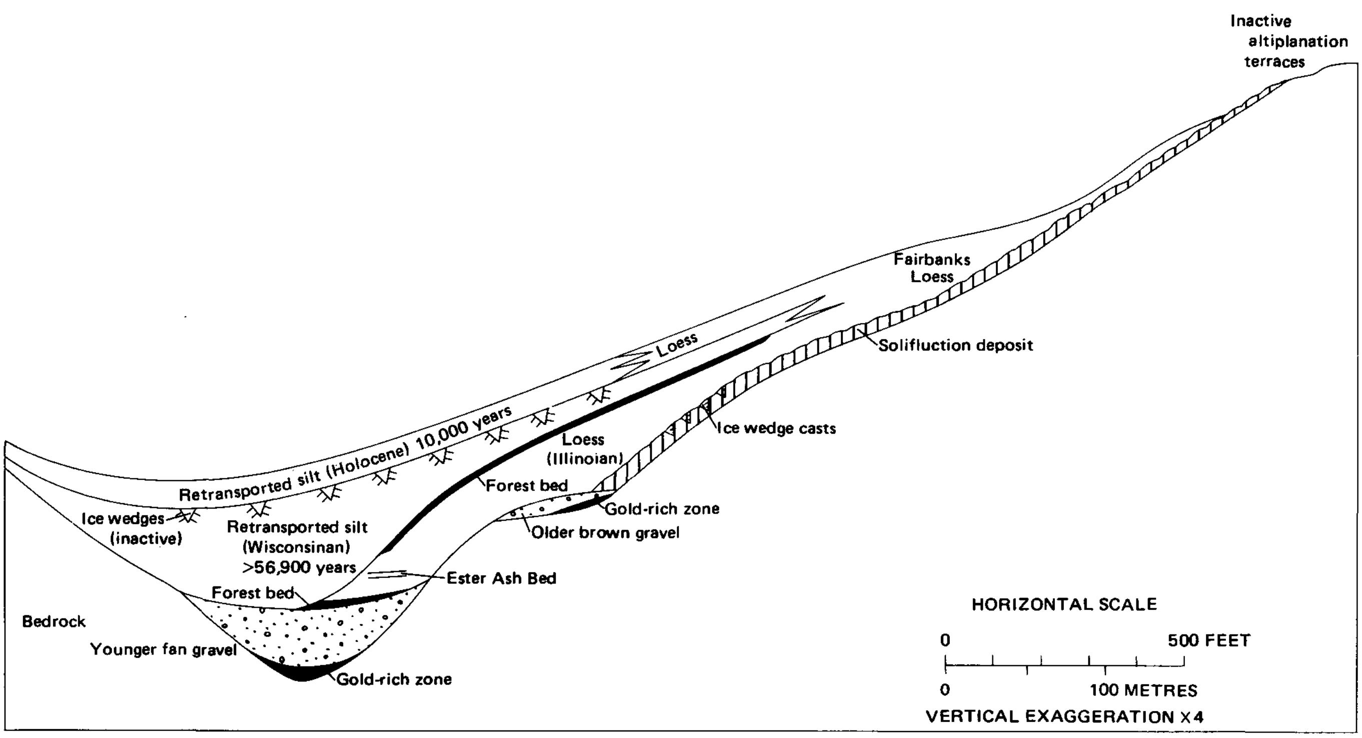


FIGURE 20.—Schematic composite cross section of a creek valley in central Alaska illustrating the relations of the Quaternary silt and gravel deposits.

TABLE 4.—*Rates of deposition of Holocene loess along rivers in central Alaska*

Area	Location	Source	Measured thickness (metres)	Duration of accumulation (years)	Rate of accumulation (mm per yr)
East side of Delta River.	Adjacent to east side of river flood plain 32–80 km south of Big Delta on terraces and alluvial fans.	Péwé, Hopkins, and Giddings (1965, p. 362).	2.4	1950±150 (L–163K)	1.14–1.33
		Reger, Péwé, Hadleigh-West, and Skarland (1964 p. 95–96).	3.8	4650±250 (L–137Q)	.78–.86
			14.0	7000±275	1.93–2.07
			5.2	5900±250 (I–646)	.85–.92
			3.7	2300±180 (I–647)	1.49–1.75
North side of Tanana River.	Adjacent to north side of river near junction with Delta River. (1) On southeast facing slope 1 km from river. (2) On low terrace at edge river.	Péwé (1965b, p. 48–49, p. 53).	(1) 0.61	2565±290 (GX0254)	.21–.27
			(2) 1.2	8040±190 (GX0255)	.15
Fairbanks	Gold Hill 70 m above and 5.4 km north of the present Tanana River.	Péwé (1965a, p. 16, 20).	6.0	4020±200 (W–183)	1.40–1.57

erosional period when most of the retransported silt in creek valley bottoms and some of the loess on the hillside slopes and hilltops were removed. The loess was deeply gullied, and block slumping occurred. Long parallel gullies more than 10 m deep and 180 m long formed on almost all loess-covered slopes in the Fairbanks area. Permafrost thawed and perhaps disappeared during this warm erosional interval (Péwé, 1975).

In Wisconsinan time, additional loess was deposited on the uplands, and much loess was retransported to valley bottoms to form a carbonaceous, fetid, perennially frozen deposit, locally termed "muck" (fig. 20). This valley-bottom facies of loess of Wisconsinan age is 3–46 m thick and contains abundant vertebrate and plant fossils, including partial carcasses of vertebrates that were entombed in the silt and perennially frozen. The most common vertebrate remains in the muck of Wisconsinan age, in order of their abundance, are those of bison, mammoth, and horse. The retransported silt (valley-bottom facies) of Wisconsinan age contains many ice wedges 0.3–3 m wide and as much as 10 m long.

The gullies and ridges of pre-Wisconsinan age cut in loess of the middle and upper slopes in the Fairbanks area were rounded and subdued by the blanket of loess deposited over these undulations in Wisconsinan time.

About 10,000 years ago a warming interval occurred that caused the permafrost table to be lowered 2–3 m and the top of the ice wedges to melt down about 0.3–3 m. Loess and the valley-bottom facies (retransported silt) of the loess that has been deposited since and during the thawing lie unconformably over the thawed-down flat-topped ice wedges and retransported silt of Wisconsinan age (fig. 20). This Holocene silt is 0.3–8 m thick, and all but the upper 1.5. m is perennially frozen (Péwé, 1975). The silt contains no bones of extinct animals.

On the hilltops, the loess deposited in Illinoian, Wisconsinan, and Holocene time constitutes one relatively uniform loess layer, and it has not yet been possible to differentiate this loess into layers of separate ages. All the upland loess is grouped together under the name Fairbanks Loess (fig. 20).

RATE OF DEPOSITION

Loess is still being deposited along most braided glacial streams in Alaska, most notably adjoining the Delta River (fig. 21) (Péwé, 1951a), along the Knik and Matanuska Rivers (Tuck, 1938; Trainer, 1961), and near the junction of the Tanana and Yukon Rivers. Information on the rate of eolian silt accumulation can be obtained in some of these areas. The 7.2 m of silt measured on the top of Bodenberg Butte in the western Matanuska Valley near Palmer (Stump and others, 1959, p. 9), for example, is entirely post-Wisconsinan in age. Measurements of loess thicknesses and radiocarbon dating of stumps and logs in the loess have yielded rates of loess accumulation in a few places in central Alaska (table 4). Assuming no erosion of the accumulating loess during the measured interval, the rate of accumulation is 0.2–2.0 mm per year. The writer considers this to be a valid figure because the areas of accumulation are flat to almost flat and heavily vegetated.

SUMMARY

Most of the low-lying areas of Alaska (except perhaps the southeastern part) are covered with either sand dunes or loess from a few millimetres to 100 m thick. Almost all these deposits are associated with glaciations. Most sand dunes are stablized and have not received detailed study.

Alaska has one of the largest, if not the largest, areas of loess in North America and perhaps contains the thickest deposits. The loess has been studied in some detail, and valuable stratigraphic sections of Illinoian, Wisconsinan, and Holocene loess are known in central and western Alaska. The sections, especially in central Alaska, illustrate a detailed history of deposition, retransportation, and permafrost and a record of mammalian remains.

The rate of loess deposition in North America has long been sought, and perhaps the best place to find the answer is in association with Holocene glaciations in Alaska. Studies here indicate loess is accumulating at 0.2–2.0 mm per year.

PERMAFROST AND PERIGLACIAL DEPOSITS

Extensive areas of Alaska are underlain by deposits that are the result of mass wasting or frost action in a

FIGURE 21.—Oblique aerial photograph of clouds of silt transported by wind from the Delta River flood plain, central Alaska. View looking north. Photograph by U.S. Navy, 1948 (Péwé, 1951a).

rigorous climate—a climate widely known as periglacial. Lozinski (1909) introduced the term "periglacial" for the concept of a cold climate outside of the glacial border; specifically, the term "periglacial" referred to the area that bordered the continental ice sheet, its climate, and features developed in this rigorous environment.

The term "periglacial" has not been accepted by all, although the concept is now firmly entrenched (Tricart and Cailleux, 1967). Such terms as "subnival," "subglacial," and "subgelid" and "frost-climate" have been suggested, but the term "periglacial" remains most widely used. No single definition agreeable to everyone studying the periglacial environment has been formulated, owing to differences of opinion concerning the climate of the periglacial zone. For instance, Zeuner (1945) restricted the periglacial zone to that area where the annual air temperature is −2° C or colder. Others stated that permafrost is not actually necessary for a periglacial environment and that such a climate could include regions where it is absent but where a considerable number of freezing and thawing cycles occur annually. Although exact definition of the periglacial environment has not been developed, workers agree that it has a rigorous climate. The writer considers all areas of existing permafrost to be in the modern periglacial zone.

The periglacial environment is characterized by frozen ground and intense frost action in fine-grained sediments and considerable sorting of fine and coarse materials. Therefore, mechanical weathering is accelerated, and ice is abundant in perennially and seasonally frozen ground. Local sorting of sediments is common, producing small-scale patterned ground. Water is concentrated in fine-grained surficial sediments by ice segregation. This concentration of water, when released, and the impeded infiltration of surface waters owing to impervious frozen subsoil aids in downslope mass movement of debris. Growth of large masses of ground ice occurs in permafrost with the formation of tundra polygons and pingos.

PERMAFROST

Permafrost, the most widespread phenomenon of the periglacial climate, is defined as a thickness of soil or other superficial deposits, or even bedrock, which has been colder than 0°C for 2 years or more; it is defined exclusively on the basis of temperature, irrespective of texture, degree of induration, water content, or lithologic character (Muller, 1945, p. 3). Permafrost, perennially frozen ground (Mozley, 1937; Taber, 1943), or *vechnaya merzlota*[1] (Sumgin, 1927) is widespread in the northern part of the northern hemisphere (fig. 22), as well as in Antarctica. About 20 percent of the land area of the world is estimated to be underlain by permafrost (Muller, 1945, pl. 1; Black, 1954, p. 842; Péwé, 1966c, 1969a, fig. 1; Ferrians and others, 1969).

Perennially frozen ground is present throughout 82 percent of Alaska (fig. 23) but is more widespread and extends to greater depths in the north than in the south (Ferrians, 1965a, b, 1966). Permafrost in Alaska has long been divided arbitrarily into three generalized zones: the continuous, discontinuous, and sporadic zones (Black, 1950, 1954; Péwé, 1954; Hopkins and others, 1955; Péwé and Paige, 1963; Wahrhaftig, 1965). The terms refer to lateral continuity of permafrost. This threefold division has followed the manner of mapping employed in the U.S.S.R., where the divisions are further qualified by the temperature of permafrost (Sumgin and others, 1940). (See Baranov (1956) for a later, more refined division.) Inasmuch as the boundaries between the discontinuous and the sporadic permafrost zones are difficult to place without temperature information, and inasmuch as thermal data for permafrost in Alaska are still limited, the writer uses only two permafrost zones in Alaska (Péwé, 1966b), the continuous and the discontinuous. A similar classification has been followed in Canada for the same reason (Jenness, 1949; R. J. E. Brown, 1960, 1966, 1969, 1970). The discontinuous zone as used here includes the discontinuous and sporadic zones of the earlier classification. In the continuous zone of the northern part of Alaska, permafrost is almost everywhere present and extends to a calculated depth of 405 m, 12 km south of Barrow (Brewer, 1958, p. 19), and to 650 m at Prudhoe Bay according to Stoneley (in Lachenbruch, 1970b, p. J3–J4). In this area, away from large bodies of water, permafrost at a depth of 15–25 m is colder than −5°C (15–25 m is the maximum depth to which appreciable annual temperature fluctuations occur).

Although permafrost was known to exist offshore northern Alaska and Yukon Territory, only recently has extensive work been undertaken to investigate its extent and origin (Shearer and others 1971; MacKay, 1972b; Lewellen, 1973). One of the pressing problems for future work is the study of offshore marine permafrost in North America.

Southward, in the discontinuous permafrost zone of Alaska, permafrost thickness decreases, and unfrozen areas are more and more abundant until, near the southern boundary (fig. 23), only rare patches of permafrost exist. The southern boundary of permafrost as shown in figure 23 is placed to include relict permafrost as well as small areas of permafrost existing because of current favorable microclimatic conditions. This boundary lies a short distance south of the 0°C mean annual

[1] This term means "eternal frost" and is being replaced in the recent Russian literature by "*mnogoletnëmerzlyy grunt*" which means "perennially frozen ground."

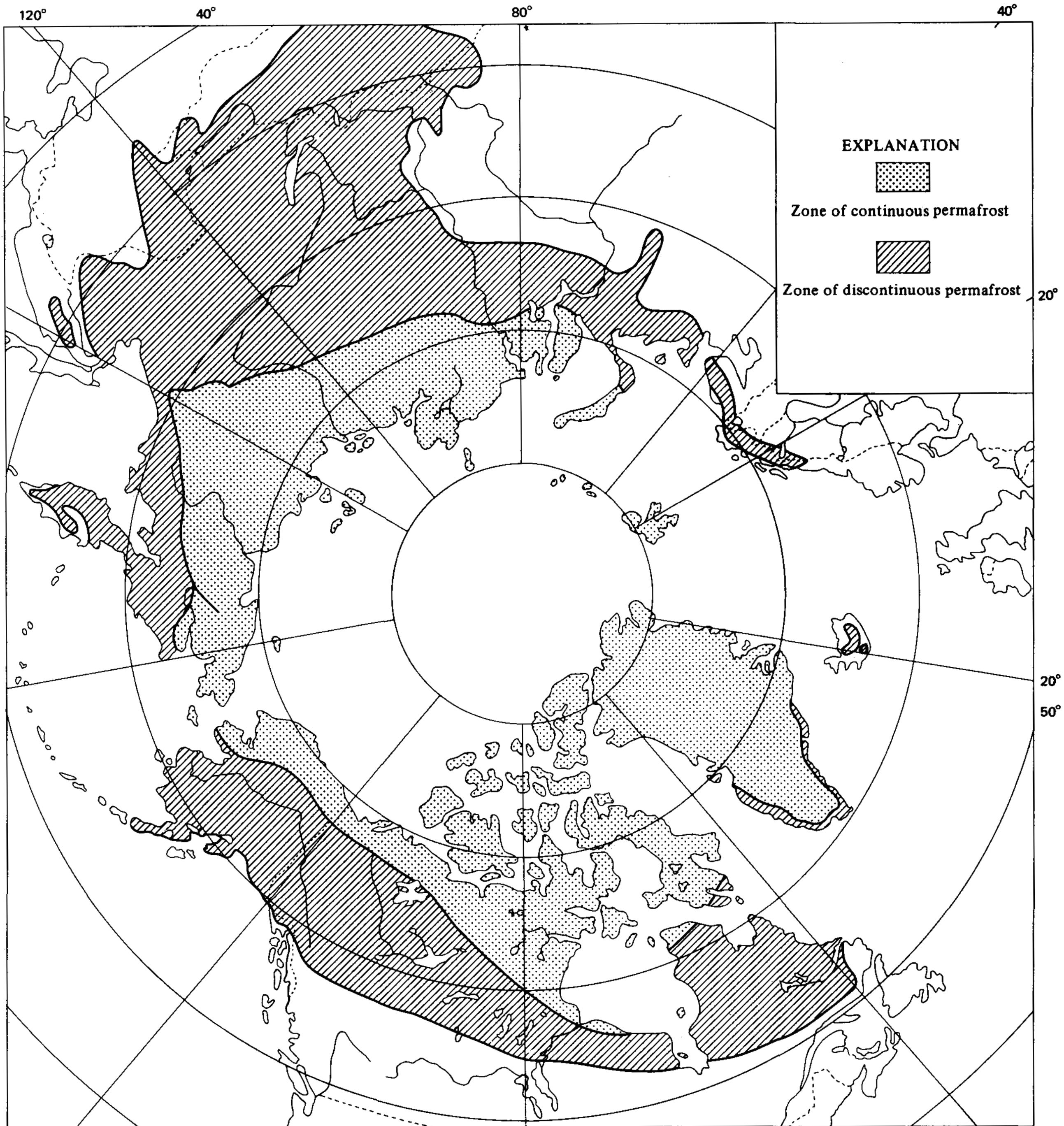


FIGURE 22.—Distribution of permafrost in the northern hemisphere. Sources: Alaska, Péwé (1969a); Canada, Brown (1969); Greenland, Weidick (1968); Iceland, Thorleifur Einarsson, Iceland Industrial Research and Development Institute, written commun., 1966; Norway, B. J. Andersen, University of Bergen, written commun., 1966; H. Svensson, Lund University, oral commun., 1966; Sweden, Rapp and Annersten (1969); Mongolia and China, Shumskiy, Schvetzov, and Dastavolov (1955); U.S.S.R., Baranov (1956). From Péwé (1969a).

air[2] isotherm, an isotherm generally thought to be now at the southern boundary of permafrost. The temperature of permafrost at a depth of 15–25 m in this zone ranges from −5°C in the northern part of the zone to approximately 0°C farther south.

[2]In this report the air temperature referred to, unless otherwise stated, is that measured at a height of 1.5 m above the ground, the standard height for the U.S. Weather Bureau recording stations.

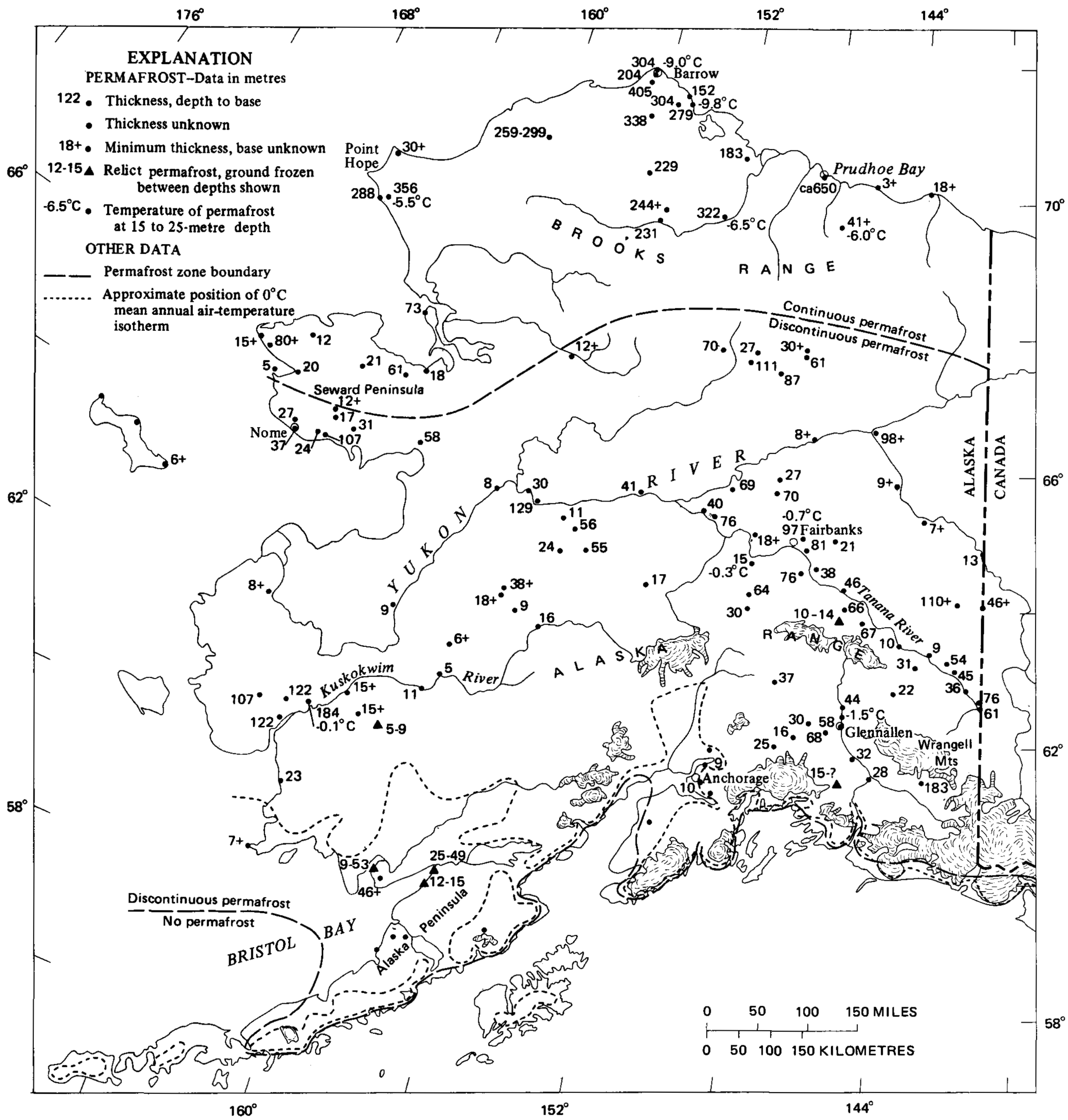


FIGURE 23.—Permafrost map of Alaska. Compiled by T. L. Péwé and L. R. Mayo. Permafrost thickness from map compiled by J. R. Williams, published and unpublished sources. Ground temperature at Glenallen measured at a depth of 6 m.

The maximum thickness of permafrost in Alaska ranges from an estimated 650 m in the north to a few metres on the Alaska Peninsula in the south (fig. 23). Permafrost has not been found in the Aleutian Islands or in southeastern Alaska. North of the Brooks Range, thickness ranges from 180 to 650 m, on the basis of a few deep test holes drilled for petroleum investigations. At Prudhoe Bay on the north coast of Alaska, holes drilled for petroleum penetrated permafrost to depths of about 650 m (Stoneley, 1970; Howitt and Clegg, 1970). Lachenbruch (1970b) noted that this is 50 percent deeper than at Barrow (fig. 23) and 100 percent deeper than at Cape Simpson, although all three localities probably have similar mean surface temperatures. He explains this difference as the result of "an increased proportion of siliceous sediments from Simpson to Barrow to Prudhoe, which would cause a corresponding increase in conductivity and decrease in geothermal gradient." Maximum thickness of permafrost in central and southwest Alaska ranges from a few metres to 180 m. Borings are numerous but mostly limited to valley bottoms and lowlands.

No attempt has been made to map isopachs of permafrost thickness in Alaska because of the almost complete lack of information in the mountainous areas. This lack of information precludes any but the broadest generalizations at this time. Bateman and McLaughlin (1920) record permafrost in the Kennecott mine in the Wrangell Mountains. This record, plus one from Keno Hill (Wernecke, 1932) at an elevation of 1,500 m, 150 km east of Dawson in adjacent Yukon Territory, Canada, constitute about the only information of permafrost at great depth in mountainous areas in Alaska and adjacent Canada.

The temperature of permafrost is one of the most critical quantitative parameters necessary to evaluate not only the existence of frozen ground at depth, but also to calculate the thicknesses, mean annual air temperature, and even history of permafrost and geomorphic changes on the surface (Lachenbruch, 1962). Accurate mapping of permafrost is possible only with abundant temperature data. Temperature data in northern Alaska (fig. 23) were collected by the U.S. Geological Survey at Cape Simpson (fig. 24), Barrow, the Shaviovik River area (Brewer, 1958), and at Ogotoruk Creek near Point Hope (Lachenbruch and others, 1961). Lachenbruch (oral commun., 1962) provided data on the permafrost at Umiat. The coldest permafrost temperature recorded in Alaska is −10.6°C, 12 km south of Barrow (Brewer, 1958, p. 20).

Permafrost temperature in central Alaska near Fairbanks is −0.9°C at a depth of 15–25 m in frozen silt (Péwé and Paige, 1963, p. 376). In the Copper River Basin the temperature of permafrost at a depth of 6 m in undisturbed ground along the Richardson Highway is −1.5°C (Greene and others, 1960, p. B141; Williams, 1970, fig. 10).

Permafrost forms when more heat leaves the ground than enters and when a temperature below 0°C is produced continuously for 2 years or more. This heat balance is delicate, and the "cold reserve" grows or shrinks as heat flow is modified by either climatic changes or by changes at the interface between permafrost and the atmosphere—that is, changes in the vegetation, snow, and characteristics of the upper layer of thawed ground.

Inasmuch as permafrost is defined on the basis of temperature alone, any material, whether it is silt, sand, gravel, peat, refuse piles, or bedrock, as long as it has been colder than 0°C continuously for more than 2 years, is termed permafrost. In some places, permafrost

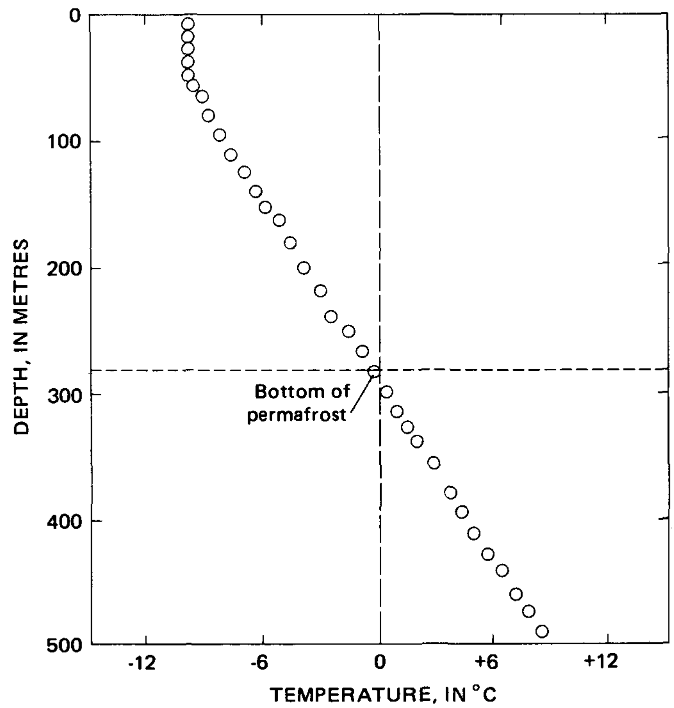


FIGURE 24.—Geothermal profile for Cape Simpson well 28, northern Alaska; data for May 22, 1953, after well had essentially returned to thermal equilibrium. Modified from Brewer (1958, p. 20).

has been formed by artificial freezing, as under cold-storage rooms in England (Cooling and Ward, 1944; Ward and Sewell, 1950) and in Canada (Hardy and D'Appolonia, 1946). Oddly enough, perenially frozen ground was created by artificial means in unfrozen loess to prevent landsliding in placer mining operations near Fairbanks, Alaska (J. D. Crawford, oral commun., Feb. 6, 1955).

Lane (1946, p. 401–402) stated that the temperature of permafrost below the depth of measureable seasonable change is approximately the same as the local mean annual air temperature if the permafrost is in equilibrium with the existing climate. Redozubov (1946, p. 2) stated that the mean annual air temperature is always slightly lower than the local permafrost temperature, in some places several degrees lower (Brewer, 1958, p. 22). Actually, at a depth where the seasonal temperature fluctuations are minimal, the average temperature of permafrost is close to the average temperature of the ground surface; however, it is 2–6°C warmer than the mean annual air temperature determined from U.S. Weather Bureau recording stations (Lachenbruch, 1970b). Temperature records of the U.S. Weather Bureau are the only ones readily available over widespread areas and therefore must be used in constructing maps of climatological data such as figure 3. The southern boundary of permafrost roughly approximates the position of the 0°C mean annual air isotherm. This supports the suggestion that most permafrost is a product of the present climate. Redozubov (1946, p. 10) showed that it would take many thousands of years to form 700 m of permafrost. Permafrost no longer in contact with seasonal frost may be a holdover from earlier colder climates.

Many temperature profiles show that permafrost is not in equilibrium with the present climate at the sites of measurements (Lachenbruch and others, 1962; Lachenbruch and Marshall, 1969; Lachenbruch, 1968, p. 835). Lachenbruch, Greene, and Marshall (1961, 1966) stated, for example, that only 260 m of the 356 m of permafrost observed in a drill hole in Ogotoruk Valley, 35 km southeast of Point Hope (pl. 1), would exist if present surface conditions were to persist for several thousands of years. "In a sense, about 25 percent of the permafrost beneath Ogotoruk Valley is a product of an extinct climate" (Lachenbruch and others, 1966, p. 158).

In a hypothetical example in central Alaska near Fairbanks, the temperature of permafrost below the zone of annual change is −1°C (fig. 25). By extrapolation downward along the known curve of the geothermal gradient, it is possible to calculate that the thickness of permafrost is about 60 m. This agrees with known thicknesses measured in the Fairbanks area. The mean annual air temperature in the region is −3°C, 2.5°C colder than the temperature of permafrost.

As the climate becomes colder or warmer (mean annual air temperature remaining below 0°C), the temperature and thickness of the permafrost change accordingly. The base level of permafrost fluctuates but not the top (the permafrost table), providing that the active layer, vegetation, and snowfall do not change. If the mean annual air temperature rises above 0°C, the top of the permafrost will be lowered by thawing. The rate at which the base or top of permafrost changes depends not only on the amount of climatic fluctuation, but also on the amount of ice in the ground. Frozen silt generally thaws more slowly than gravel because it generally contains more ground ice. Many unfrozen areas in the discontinuous zone of permafrost in Alaska (fig. 23) consist of sand and gravel, and the nearby perennially frozen areas consist of silt (Péwé and others, 1969).

GROUND ICE

Ice content is probably the most important feature of permafrost affecting human activity in the north. Of equal importance is the fact that it provides evidence concerning past climates beyond the range possible by geothermal calculations. Ice in the perennially frozen ground varies in size and shape and has definite distribution characteristics. The origin of ground ice constitutes one of the most interesting and controversial problems connected with permafrost. The writer groups

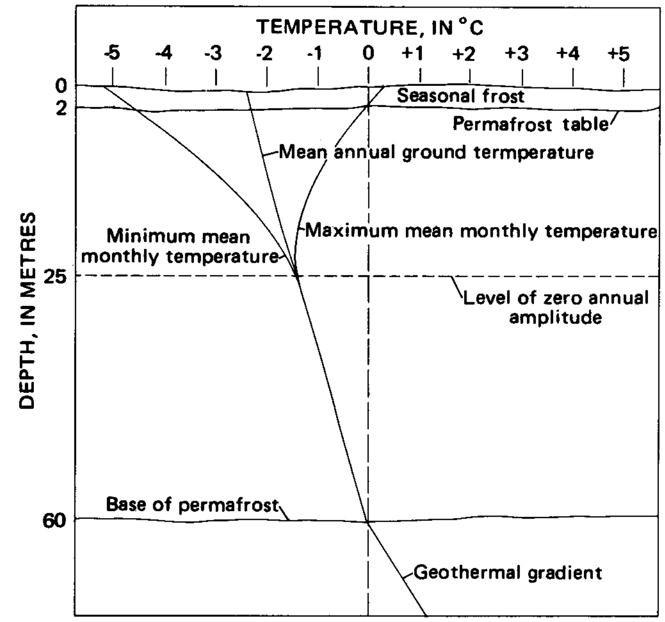


FIGURE 25.—Hypothetical example showing temperature profile and thickness of permafrost in central Alaska.

ground ice into five main types: (1) pore ice, (2) segregated or Taber ice, (3) foliated or ice wedge ice, (4) pingo ice, and (5) buried ice.

Pore ice is defined as ice that fills or partly fills pore spaces in the ground. It forms by freezing pore water in situ with no addition of water. The ground contains no more water in the solid state than it could hold if the water were in the liquid state. Black (1954) termed frozen ground with such ice as undersaturated or saturated.

Segregated or Taber ice is described as ice seams, lenses, or layers generally 1–100 mm thick that grew in the ground by drawing in water as the ground became frozen. Taber (1929, 1930) was a pioneer in demonstrating this phenomenon, although Beskow (1947) worked actively in this field at about the same time. Although the idea that water must be supplied for an ice crystal to grow is generally accepted, not everyone agrees on how this is accomplished (Jackson and Chalmers, 1956; Jumikis, 1955; Cass and Miller, 1959; Winterkorn, 1943).

Segregated ice has been referred to by various terms—ice seams, ice segregations, ice gneiss (Taber, 1943), and sirloin-type ice (Higashi, 1958), among others. The writer has long used the term "Taber" ice, originally suggested to him by A. H. Lachenbruch, U.S. Geological Survey, for ice segregations in the ground.

Pore ice and Taber ice occur both in seasonally frozen ground and in permafrost. Black (1954) referred to permafrost composed of this type of ice as being supersaturated because it contains more water in the solid state than the ground could possibly hold if the water were in the liquid state.

Foliated ground ice or wedge ice is the term given to large masses of ice that grow in thermal contraction cracks in permafrost.

Pingo ice is clear or relatively clear ice that occurs in permafrost in nearly horizontal or lens-shaped masses 3–15 m in diameter. It evidently originates from ground water under hydrostatic pressure (Müller, 1963).

Buried ice in permafrost includes buried sea, lake, and river ice and recrystallized snow. Buried blocks of glacier ice in a permafrost climate would also fall into this category.

Small ice segregations are the least spectacular and yet one of the most extensive types of ground ice. They occur in both seasonally and perennially frozen ground, and as grains, granules, films, lenses, layers, pods, dikes, and irregular masses that range in diameter from less than 1 mm to 4 cm. The freezing of ground and formation of small, clear ice segregations has received much study from engineers and geologists interested in ice growth and its effect on engineering structures. As population increases in the far north, the effects of the growth of ground ice on construction have been the subject of growing concern (Beskow, 1947; U.S. Army Corps of Engineers, 1946, 1947, 1949, 1954, 1956; Johnson, 1952; Reed, 1969).

When considering the subject of ground ice, the question if generally raised as to how much ice actually exists in the ground. Because such information would be interesting and valuable from a historical standpoint as well as essential in solving engineering problems posed by permafrost (Lachenbruch, 1970a), the question is being considered in an ambitious attempt to make an inventory of all perennial ice and snow masses on and beneath land surfaces of the earth as a project of the International Hydrological Decade (Fritz Müller, unpub. data, 1968).

Estimates the volume of worldwide ground ice range from 0.2 to 0.5 million km^3, less than 1 percent of the total volume of the earth (Shumskiy and Vtyurin, 1966; Shumskiy and others, 1964, p. 433). Estimates of ground ice in Alaska are sorely needed and represent a challenging project. Brown (1967c) made a first approximation of the ground ice inventory for the Arctic Coastal Plain (pl. 1). The plain lies in the continuous permafrost zone (fig. 23) and represents a fairly uniform set of geologic, geomorphic, and permafrost conditions. An underground ice inventory for other regions of Alaska would be more difficult owing to discontinuities in the permafrost and the presence of large ice masses with no physiographic expression.

On the basis of an examination of ice in the ground in many bore holes in the Barrow area (Brown, 1965b, 1966a, b, 1967b, 1969a, b; Brown and Johnson, 1965; Sellmann and Brown, 1963, 1965; Sellmann and others, 1964, 1965) and the extrapolation of information from the bore holes to the rest of the coastal plain by use of aerial photographs and geologic maps, it is estimated (Brown, 1967c) that 10 percent by volume of the upper 3.5 m of permafrost on the coastal plain is composed of ice wedges (foliated ground ice); Taber ice is the most extensive type of ground ice, in places representing 75 percent of the ground by volume. In measurements made by the writer in the Fairbanks area, samples of frozen ground may contain as much as 200–1,000 percent ice by weight (Péwé, 1958b; Williams and others, 1959). For purposes of computation, Brown (1967c) assumed that no segregated ice exists below 8 m and that the only ground ice present is pore ice. This is probably correct for most areas, but especially in syngenetic ice areas, the writer has seen deposits below a depth of 8 m with ice seams 1 cm thick and buried ice 2 m thick. Brown calculated that the pore and Taber ice content in the depth between 0.5 and 3.5 m (surface to 0.5 m is seasonally thawed) is 61 percent and between 3.5. and 8.5 m is 41 percent by volume. The total amount of pingo ice is less than 0.1 percent of the permafrost. The total amount of perennial ice in the permafrost of the Arctic Coastal Plain is estimated to be 1,500 km^3 and below 8.5 m most of that is present as pore ice.

ICE WEDGES

The most conspicuous and controversial type of ground ice in permafrost is the large ice wedges or masses characterized by parallel or subparallel foliation structures. Most foliated ice masses occur as wedge-shaped, vertical, or inclined sheets or dikes 1 cm to 3 m wide and 1–10 m high where seen in transverse cross section (fig. 26). Some masses seen on the face of frozen cliffs may appear as horizontal bodies a few centimetres to 3 m in thickness and 0.5–15 m long. The true shape of these ice wedges can be seen only in three dimensions. Ice wedges are parts of a polygonal network of ice enclosing polygons or cells of frozen ground 3–30 m or more in diameter.

The network of foliated ice in the ground generally causes a microrelief pattern on the surface of the ground, generally called polygonal ground or tundra polygons (fig. 27). Troughs that delineate polygons are generally underlain by ice wedges 1–2 m wide at the top. These polygons are 2–30 m in diameter and should not be confused with small-scale polygons or patterned ground produced by frost sorting.

Polygons may be low centered or high centered. Upturning of strata adjacent to the ice wedge may make a ridge of ground at the surface on each side of the wedge (Leffingwell, 1919, fig. 25 and pl. 29b), thus enclos-

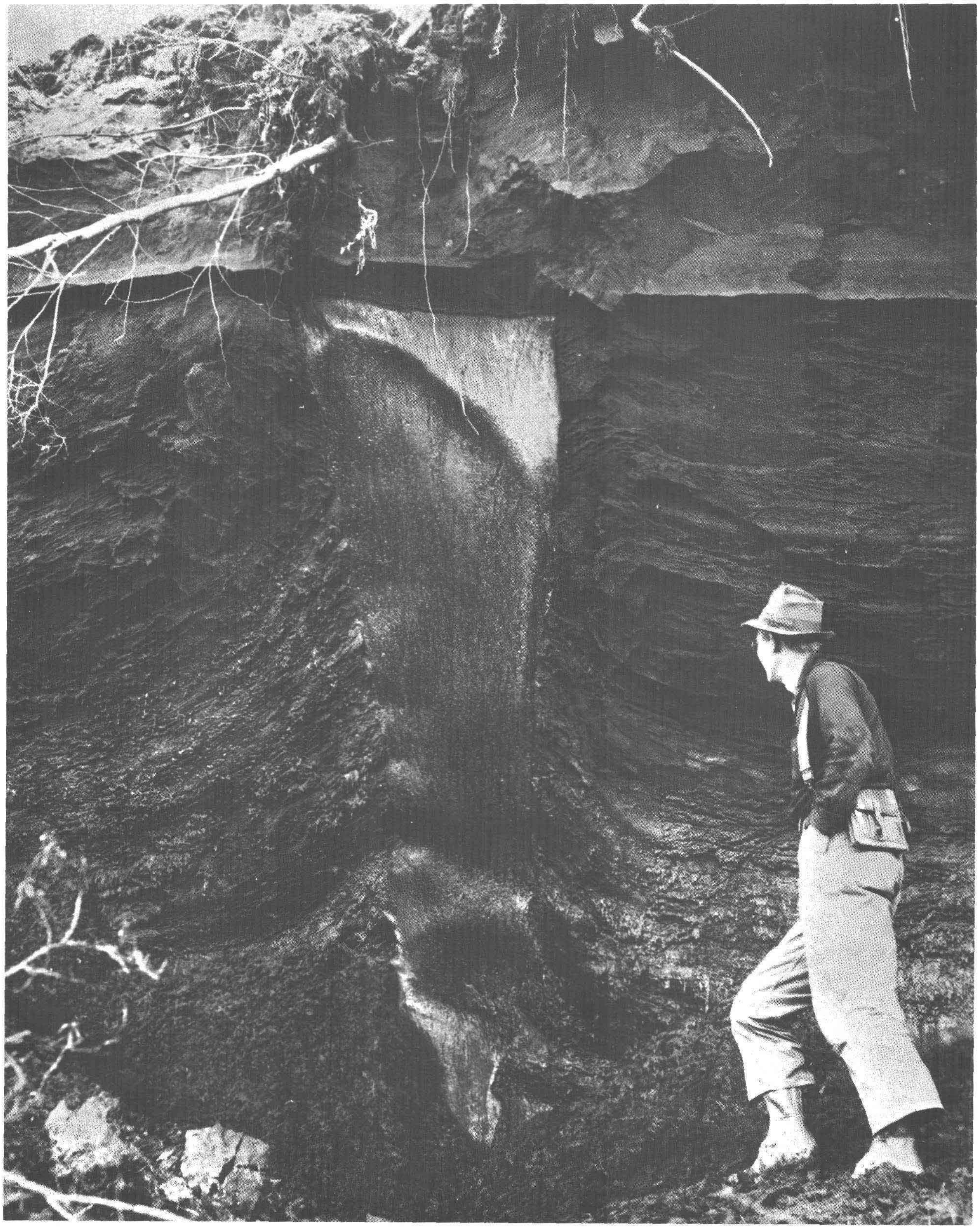


FIGURE 26.—Foliated ground-ice mass (ice wedge) in silt exposed in placer gold mining operations on Wilber Creek near Livengood, Alaska. Photograph No. 474 by T. L. Péwé, September 19, 1949.

FIGURE 27.—Oblique summer aerial view of raised-edge ice wedge polygons on the northern Alaskan sea coast near Barrow, Alaska. Diameters of polygons are 7–15 m. Photograph by R. I. Lewellen, August 11, 1966.

ing the polygons. Such polygons are lower in the center and are called low-centered polygons or raised-edge polygons. They are erroneously termed by some "depressed center" polygons. Low-center or raised-edge polygons indicate that ice wedges are actually growing and that sediments are being actively upturned (Brown, 1969c, p. 124). If erosion, deposition, or thawing is more prevalent than the up-pushing of the sediments along the sides of the wedge, or, if the material being pushed up cannot maintain itself above ground level, low ridges will be absent, and there may be either no polygons at the surface or the polygons may be higher in the center than troughs over the ice wedges that enclose them. Such high-center polygons are erroneously termed "raised center" polygons by some. Both high-center and low-center tundra polygons are widespread in Alaska and are good indicators of the presence of foliated ice masses; but care must be taken to demonstrate that the pattern is not a relic and evidence of ice wedge casts.

The origin of large ground-ice masses in perennially frozen ground of North America has been discussed in print since Kotzebue recorded ground ice at Elephant Point in Eschscholtz Bay (Kotzebue, 1821). The origin of ground ice was first discussed in Siberia (Bunge, 1884, 1902). The general theory for the origin of ice wedges now accepted is the thermal contraction theory of Leffingwell (1915, 1919), which was succinctly summarized by Lachenbruch (1962):

> During the Arctic winter, vertical fractures on the order of one-tenth of an inch wide and several feet deep are known to form in the frozen tundra—this process is generally accompanied by loud reports. They are assumed to be the result of tension caused by thermal contraction of the tundra. In early spring it is supposed that water from the melting snow freezes in these cracks and, with accumulating hoarfrost, produces a vertical vein of ice that penetrates permafrost. Horizontal compression caused by reexpansion of the permafrost during the following summer results in the upturning of permafrost by plastic deformation. In the winter that follows, renewed thermal tension supposedly reopens the vertical ice-cemented crack which is presumed to be a zone of weakness. Another increment of ice is added when the spring meltwater enters the renewed crack and freezes. Such a cycle, it is argued, acting over centuries, would produce the vertical wedge-shaped mass of ice. The polygonal configuration is generally thought to be a natural consequence of contraction origin.

The thermal contraction hypothesis receives the most support (Bunge, 1884, 1902; Leffingwell, 1915, 1919; Black, 1952b, 1954, 1963; Popov, 1955, 1965; Hopkins and others, 1955; Shumskiy and others, 1955; Washburn, 1956; Britton, 1958b; Péwé, 1958a, 1959, 1962, 1966a, b; Lachenbruch, 1961, 1962, 1966; Brown, 1967a; Journaux, 1969). Taber (1943) and Schenk (1965), however, do not agree with this hypothesis.

For the maximum annual thermal tension at the top of the permafrost to be of the same order of magnitude as the strength of foliated ground ice, the ground must cool at a certain rate for a certain period of time during a winter cold snap. A. H. Lachenbruch (oral commun., 1962) stated that although the relations that determine whether an ice wedge cracks are extremely complex, a single simple criterion that takes account of many of the factors is the minimum winter temperature at the top of the permafrost. He suggested that when its value is below −15° to −20°C, the active cracking of ice wedges might be expected in many permafrost materials.

Ice wedges may be classified as (1) active, (2) inactive, and (3) ice wedge casts. There is a complete gradation from one category to the next in areal distribution. Active ice wedges are defined as those which are actively growing. The wedge may not crack every year, but during many or most years cracking does occur, and an increment of ice is added. Under these conditions, low-center (raised-edge) polygons are well developed and ubiquitous. Such microtopography is widespread in polar areas of actively growing ice wedges and sand wedges (Péwé, 1959; Péwé and Church, 1962). High-center polygons may also be present.

The area of active ice wedges in Alaska appears to coincide roughly with the continuous permafrost zone (fig. 28) and is, in a general way, restricted to northern and northwestern Alaska. The meager thermal data available indicate that the winter temperature of the ground at the top of permafrost is about −15°C or colder. This area is limited almost entirely to tundra. Here active ice wedges occur in silt, sand, and gravel. From north to south across Alaska, a decreasing number of wedges crack frequently. The line dividing zones of active and inactive ice wedges is arbitrarily placed at the position where low-center or raised-edge polygons are uncommon and where it is thought most wedges do not frequently crack. When more data become available concerning the termperature at the top of the permafrost, this arbitrary line may be more accurately placed.

The area of active wedges in Alaska outlined in figure 28 has the most rigorous climate of the State. The mean annual air temperature ranges from about −6° to −8°C on the south to −12°C at Barrow on the north (fig. 3). Minimum temperature of the ground in winter at the permafrost table (top of the ice wedge) ranges from about −15°C on the south and −11° to −14°C near Kotzebue (estimated from incomplete data from Cold Regions Research and Engineering Laboratory (CRREL), U.S. Army Corps of Engineers, Hanover, N.H., 1963) to about −30°C in the far north near Barrow (Lachenbruch, 1962). Temperature at the top of the permafrost table at Ogotoruk Creek near Point Hope is about −18° to −20°C on the basis of a two-year record (A. H. Lachenbruch, oral commun., Nov. 20, 1962).

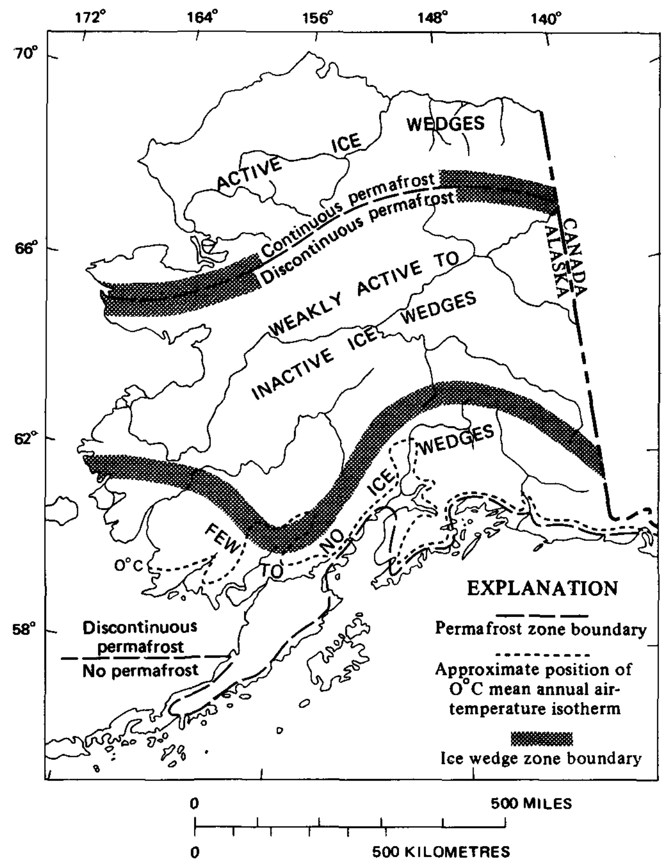


FIGURE 28.—Distribution of ice wedges and permafrost in Alaska. From Péwé (1966b).

Inactive ice wedges are defined as those that are no longer growing. The wedge does not crack in winter, and therefore no new ice is added. Of course, a gradation between active ice wedges and inactive ice wedges is represented by those wedges that crack rarely. Inactive ice wedges have no ice seam or crack extending from the wedge upward to the surface in the spring. The wedge top may be flat (fig. 26), especially if thawing has lowered the upper surface of the wedge at some time in the past. Low-center or raised-edge polygons are absent or rare, but high-center polygons are common.

The area of inactive ice wedges in Alaska lies south of the Brooks Range and the south half of the Seward Peninsula (fig. 28); the northern boundary of the zone roughly coincides with the discontinuous-continuous permafrost line (fig. 23). This region extends south to the Alaska Range and almost to the lower Kuskokwim drainage. Few data are available to permit the placing of even a broad line indicating the southern border of inactive ice wedges in the Yukon–Kuskowim Delta area. Inactive ice wedges in Alaska have been described only from frozen silt (Taber, 1943; Péwé, 1948, 1958b, 1962, 1966b); none are known to exist entirely in sand or gravel. The inactive ice wedges are of Wisconsinan age.

In the area of inactive ice wedges outlined in figure 28, the mean annual air temperature ranges from about −2°C in the south to about −6° to −8°C in the north (fig 3). Minimum winter temperatures at the top of permafrost are −3.3°C (temperature data for the cold winter of 1961–62 from F. Kitze, oral commun., 1963) near Fairbanks, −4°C at Northway (temperature data from CRREL, U.S. Army Corps of Engineers, Hanover, N.H., 1963), are estimated to be −3° to −6°C in the Copper River Basin on the basis of a 4–6-year record (A. H. Lachenbruch, oral commun., Nov. 20, 1962), and are suggested to be −10° to −15°C near the northern border of the zone. Such ground temperatures probably rarely permit thermal cracking of the ice wedges; therefore, no or rare ice increments are added to existing ice wedges, and they can be considered dormant, relict, or inactive.

In this zone the permafrost in some areas of permeable sand and gravel has been thawed by heat supplied from moving ground water. Ice wedges that formerly existed in such sediments, therefore, are no longer present (Péwé and others, 1969).

South of the zone of inactive ice wedges in Alaska, there lies a zone of discontinuous permafrost that contains practically no ice wedges. This area lies south of the Alaska Range and includes the Copper River Basin, the middle Susitna River valley, the Bristol Bay lowland (E. H. Muller, written commun., Mar. 8, 1963), and perhaps the southern part of the Yukon–Kuskokwim Delta (fig. 28). Ice wedges are not growing in this zone now. Perhaps they did not form in most of this zone during the Wisconsinan Glaciation because during all or part of Wisconsinan time most of the zone was under massive glaciers or proglacial lakes. After withdrawal of the ice or lakes, permafrost has formed locally, but the climate evidently has not been rigorous enough for many ice wedges to form except perhaps for a few small wedges that have grown in favorable places where the wind blows the snow cover away or where the vegetation cover is thin. Nichols reported (1966; written commun., 1968) that a few small ice wedges of Holocene age exist in the Copper River Basin. Areas outside glacial advances of late Wisconsinan age may have had ice wedges, but many or all of these wedges have now melted, as exemplified by the presence of ice wedge cast polygons in the Bristol Bay lowland.

AGE OF ICE WEDGES

The writer examined ice wedges in Alaska, Canada, Scandinavia, Siberia, and Antarctica and considers all wedges he has seen as Wisconsinan or Holocene in age. None to his knowledge have been proved to be pre-Wisconsinan. Wedges in Alaska are of several ages, but none appear older than the last major cold period—the Wisconsinan. Those formed prior to that time have either melted or are so rare or deeply buried that they are no longer exposed or penetrated in drilling.

Most wedges are dated by geologic association; however, attempts have recently been made to date wedges by radiocarbon analyses of the organic debris in the ice. This debris is primarily washed down the contraction crack throughout its history or is incorporated from the sides as the wedge grows. Radiocarbon dates should be younger or no older than the enclosing sediments, and this is borne out by samples collected by the writer (fig. 29). A wedge at Barrow was dated at 14,000 years old (Brown, 1966b), wedges at Fairbanks near Fox at 31,400 and 32,300 years old (Sellmann, 1967, 1972), and at Eva Creek (fig. 29) and Ready Bullion Creek near Fairbanks as older than 25,000 years. Problems of dating ice wedges using radiocarbon have not yet been solved. One of the first refinements probably should be to date samples collected systematically along vertical and horizontal lines from a "typical" large wedge. Associated soil and wood should also be dated as illustrated in figure 29. Another method to study the age of ice wedges would involve the use of the O^{16}–O^{18} ratio in the ice. To the writer's knowledge, this technique has not yet been tried.

ICE WEDGE CASTS

In places in western, central, and south-central Alaska ice wedges have melted, and these voids have been filled with sediments collapsing from above and the sides. Although it is well known that ice wedges are common in Alaska, it is not generally realized that ice wedge casts exist. Actually, although most work on ice wedge casts has been done in Europe (see Péwé and others, 1969, for list of references; see Brown and Péwé, 1973, and Péwé, 1973, for discussion of ice wedge casts and their paleoclimatic significance in North America), Alaska offers perhaps a great opportunity for the study and use of ice wedge casts as paleoclimatic indicators. This is because a good record of pre-Illinoian, Illinoian, and Wisconsinan permafrost exists in Alaska as well as sediment replacements of ice wedges of those ages. Such is not recorded elsewhere. Also, in Alaska, ice wedges and permafrost of Wisconsinan age still exist and aid

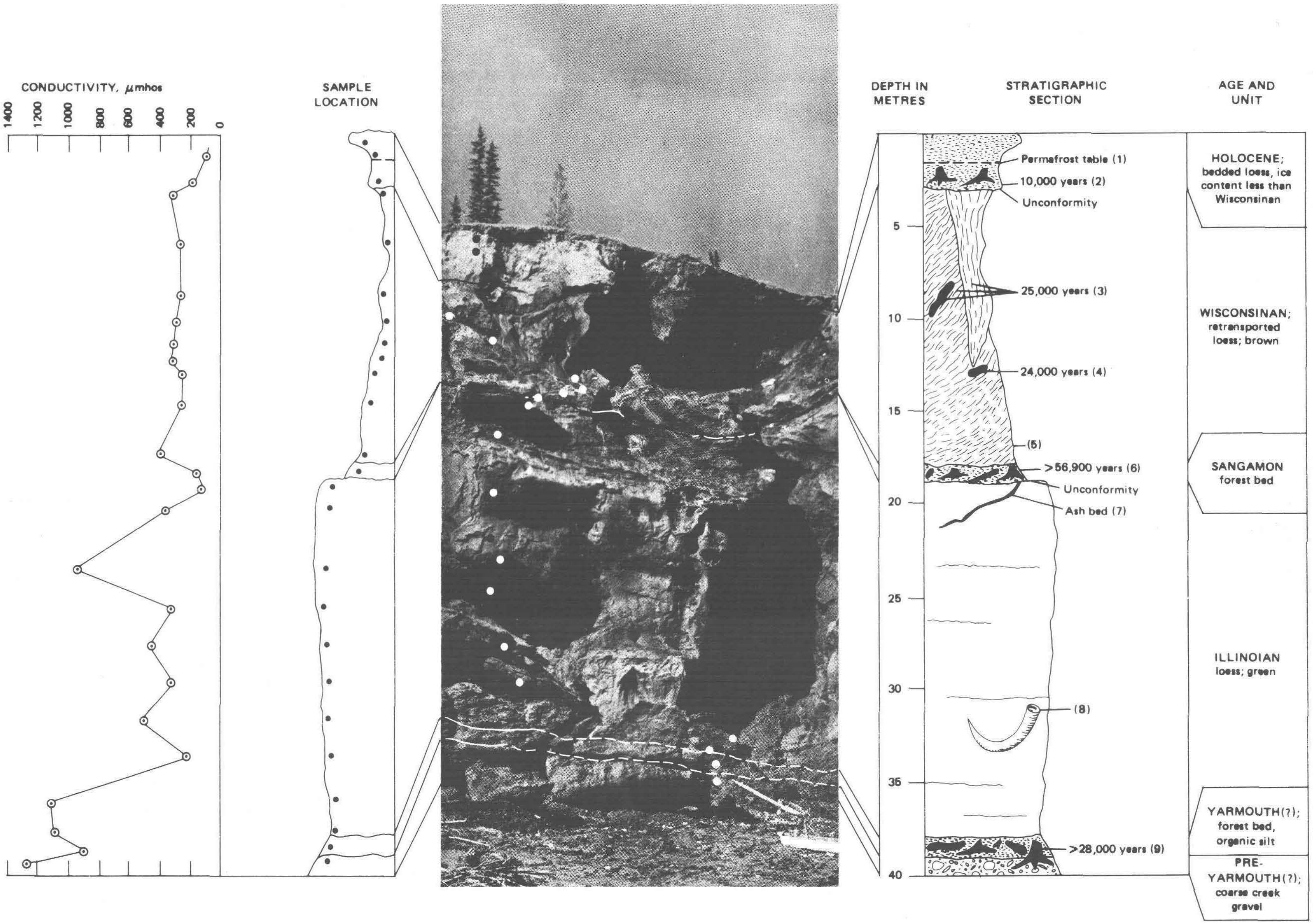


FIGURE 29.—Stratigraphic section of late Quaternary perennially frozen sediments with radiocarbon dates, vertebrate occurrences, and geochemical analyses of sediment. Eva Creek, 16 km west of Fairbanks, Alaska. Photograph No. 2540, by T. L. Péwé, July 2, 1967. See table 5.

Notes for figure 29:

(1) The permafrost table was lowered throughout central Alaska 1–5 m about 10,000 years ago and has subsequently risen to its present position. All sediments in the deep thaw layer were unfrozen for hundreds if not thousands of years. Flat-top ice wedges and collapse of sediments over ice wedges remain as evidence of this thawing episode.

(2) The age of this formation in the Fairbanks area is well documented (Péwé, 1952a, 1958b, 1965a, 1975). More than a dozen samples have been dated from the Holocene loess and retransported loess in many different mining exposures; the dates range from about 3,000 to 10,000 years, with most of them in the neighborhood of 5,000–7,000 years. A date of 3,750±200 years (L–117H) was obtained on a sample of wood collected by the writer in 1951 from the Holocene loess and retransported loess in Eva Creek 2 m below the surface. A date of 10,000±500 years (L–137S) was obtained on a sample of wood collected by the writer in 1951 from the base of the Holocene loess on Eva Bench mining exposure 3 m below the surface.

(3) In 1967 the writer collected for dating samples of ice, wood, and organic-rich silt that were adjacent. All were dated by the radiocarbon laboratory at the University of Arizona as more than 25,000 years old. More detailed dating probably would have indicated an age between 25,000 and 30,000 years. A wood sample collected in 1951 by the writer 2.5 m below the top of the Wisconsinan sediments on Eva Creek was dated at 23,300±1,000 years (W–435). Wood samples collected by the writer in 1952 at the base of the Wisconsinan formation and near the middle were dated as greater than 23,000 years old (L–157A) and greater than 30,000 years (L–163J), respectively.

(4) A radiocarbon date of 24,400±650 years (I–2116) was obtained on wood collected at the base of the ice wedges (Matthews, 1968a, p. 207).

(5) Matthews (1968a, b) examined fossil beetles from Wisconsinan and Sangamon deposits. Beetles, plus a pollen study by Matthews and of samples collected by the writer, indicate a treeless tundra environment during Wisconsinan time. In 1949 Péwé also collected bones of mammoth and *Citellus undulatus* from Wisconsinan sediments here. Guthrie (1968b) recovered *Microtus gregalis, Lemmus sibericus, Dicrostonyx torquatus,* and *Citellus undulatus* from both Wisconsinan and Illinoian deposits in this section.

(6) Péwé (1952a) reported a forest bed at this stratigraphic horizon from many exposures in the Fairbanks area. The oldest date obtained on this bed was reported by Matthews (1968b) as more than 56,900 years (Hv–1328). This organic-rich silt layer with tree stumps and logs unconformably overlies the deformed loess. (See fig. 20).

(7) In innumerable exposures of the contact between Wisconsinan and Illinoian silt in the Fairbanks area a 1-cm-thick, white, vitric, volcanic ash bed occurs at the top of the Illinoian loess and is truncated by an unconformity. This bed is deformed by faulting and solifluction (Péwé, 1975).

(8) The writer collected (1949–67) mammoth tusks and various bones of bison from loess of Illinoian age of Eva Creek.

(9) At Dawson mining cut and others a well-developed organic silt crops out (Péwé, 1952a, 1975). This unit contains large stumps and logs that are partly flattened and deformed; it is thought to represent an interstadial or interglacial forest bed. One particularly fine white spruce stump (fig. 39) from the organic silt of the Eva Creek section and with roots in the underlying gravel is more than 28,000 years old (L–137X).

tremendously in the interpretation of description, origin, and paleoclimatic importance of ice wedge pseudomorphs. The presence of ice wedges in permafrost offers a refinement in paleoclimatic interpretation not possible in permafrost without ice wedges.

Although ice wedge casts have been reported from southern Canada and the northern part of the United States, they have been most thoroughly studied in North America in Alaska, particularly in glacial outwash of Wisconsinan age near Big Delta (pl. 1) (Péwé and others, 1969). To date, of the 21 localities where ice wedge casts have been mentioned in North America outside Alaska, none have been demonstrated to be pre-Wisconsinan in age. In Alaska both pre-Wisconsinan (Illinoian(?)) and pre-Illinoian ice wedge casts are present (table 5; fig. 30).

In Alaska ice wedge casts in sediments of Wisconsinan age occur south of the glacial border in Bristol Bay (Hopkins and others, 1955, p. 139), on the Pribilof Islands (D. M. Hopkins, oral commun., Dec. 4, 1972), and near Nome (Hopkins and others, 1960). Ice wedge casts indicative of Wisconsinan permafrost have been reported in gravel along the Tanana valley outside the Wisconsinan ice limits (Péwé and others, 1969; Blackwell, 1965; Ager, 1972). Ice wedge casts in the south probably indicate a northward retreat of the southern boundary of permafrost. Those in the Tanana valley indicate a thawing of permafrost in well-drained gravel areas with a minimum of vegetation and winter snow cover.

In addition to aiding in the reconstruction of the distribution of permafrost in Alaska during Wisconsinan time, ice wedge casts indicate the earlier existence of permafrost, for example in Illinoian(?) time on the Pribilof Islands (Hopkins and Einarsson, 1966) and various places in northwestern Alaska (McCulloch and others, 1965; McCulloch and Hopkins, 1966). Ice wedge casts in two areas in Alaska permitted the writer to delineate permafrost of an extremely ancient age, perhaps a million years or more. In central Alaska Péwé (1965b, fig. 5–22) described well-developed ice wedge casts of pre-Illinoian age in solifluction deposits at Shaw Creek. Also, an ice wedge casts of sand underlying two different ages of loess are exposed in excavations on the University of Alaska campus and are thought to be pre-Illinoian in age (Péwé, 1965a, p. 17).

TABLE 5—*Ice wedge cast localities in Alaska*

Location No. (fig. 30)	Author	Date	Locality	Time of ice wedge formation
1	Hopkins, MacNeil, and Leopold.	1960	Nome	Wisconsinan.
2	Péwé	(unpub. data)	Fairbanks	Wisconsinan.
3	Blackwell	1965	Tanana River	Wisconsinan.
4	Péwé, Church, and Andreson.	1969	Big Delta	Wisconsinan.
5	Ager	1972	Healy Lake	Wisconsinan.
6	Hopkins	1972 (oral commun.)	Pribilof Island	Wisconsinan.
7	Hopkins	1972 (oral commun.)	Kvichak Peninsula	Wisconsinan(?).
8	Hopkins and others.	1955, p. 139	Bristol Bay	Wisconsinan.
9	Hopkins and Einarsson.	1966	Pribilof Island	Illinoian(?).
10	McCulloch and Hopkins.	1966	Northwestern Alaska	Illinoian(?).
11	McCulloch, Taylor, and Rubin.	1965, p. 449	Baldwin Peninsula, northwestern Alaska.	Illinoian(?).
12	Péwé	1966b	Fairbanks	Illinoian(?).
13	Guthrie and Matthews.	1971	Northern Seward Peninsula.	Pre-Illinoian.
14	Péwé	1965a, p. 17	Fairbanks	Pre-Illinoian.
15	Péwé	1965b, fig. 4–22	Shaw Creek	Pre-Illinoian.

In the far western part of Alaska, Guthrie and Matthews (1971) indicated the presence of ice wedge casts near Cape Deceit which were thought by Hopkins (1972) to be at least a million years old.

PINGOS

Another type of massive ground ice in permafrost is that found in pingos. Pingos are conical ice-cored hills or mounds, round to oval in plan, 20–400 m in diameter, and 10–70 m high that form when large, massive layers of ice grow near the surface in permafrost (Porsild, 1938; Muller, 1963; MacKay, 1966; Holmes and others, 1968). Pingos are of two distinct genetic types: the closed system and the open system. The closed-system type forms in nearly level areas when unfrozen ground water migrates under pressure to a site where then permafrost is domed up to form a mound. These are the larger of the two types of pingos and occur in areas of continuous permafrost, the tundra areas. The open-system type is generally smaller and forms on sloping ground where water beneath, or within, the permafrost penetrates the permafrost under high hydraulic pressure, which along with crystallization pressure heaves the overlying material to form a mound.

The distribution of pingos in Alaska has become much better known in the last decade. The presence of pingos are not only of value in indicating the presence of permafrost, but in indicating continuous or discontinuous permafrost. With one or two exceptions, the first local maps of pingo distribution in Alaska and adjacent Canada have appeared in the last 10 years (MacKay, 1963, 1973; Holmes and others, 1968; Hughes, 1969); the first map of pingo distribution in north-northwest North America is shown in figure 31.

Considerable progress has been made in the discovery and mapping of many open-system pingos in unglaciated central Alaska and adjacent Yukon Territory (fig. 31), as well as in the discovery of pingolike mounds in the shallow waters of the Beaufort Sea north of the mouth of the Mackenzie River (Shearer and others, 1971). The greatest advance in pingo research in the last decade has been a consideration and understanding of theory and rate of pingo growth by MacKay (1968,

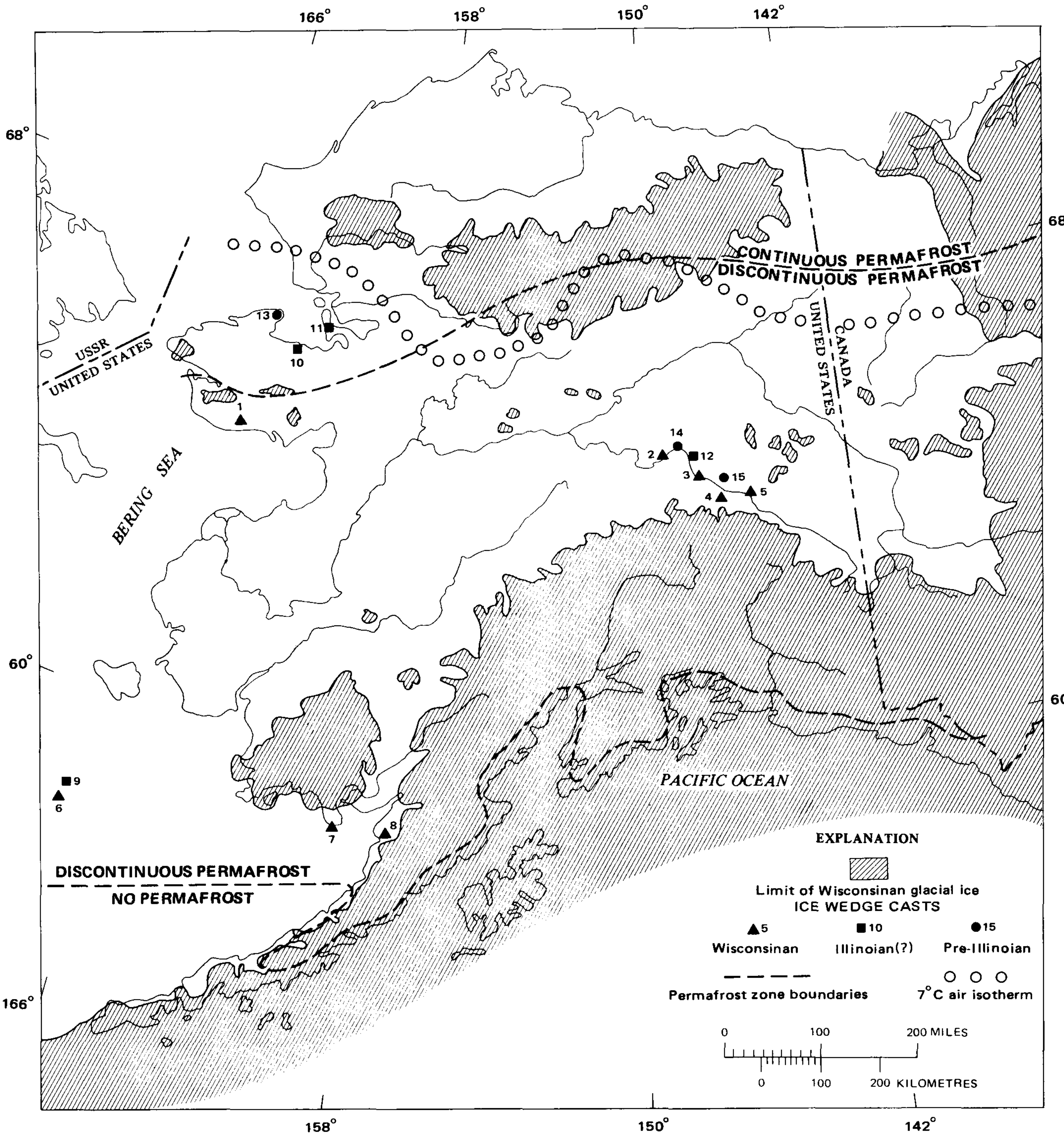


FIGURE 30—Distribution and age of ice wedge casts in Alaska.

1971, 1972a, b, 1973; MacKay and Stager, 1966; MacKay and Black, 1973).

After restudy of the photographs of the mounds in the Yukon-Kuskokwim Delta in Alaska and an examination of palsas elsewhere in the world, Péwé believes that the 200 so-called closed-system pingos mentioned by Burns (1964) occurring on the Yukon-Kuskokwim Delta are probably not pingos, but palsas.

Pingos are abundant in the forested, subarctic, discontinuous permafrost zone of Alaska. Although Harrington photographed pingos in central Alaska 60 years ago (Leffingwell, 1919, pl. 17A; Mertie and Harrington,

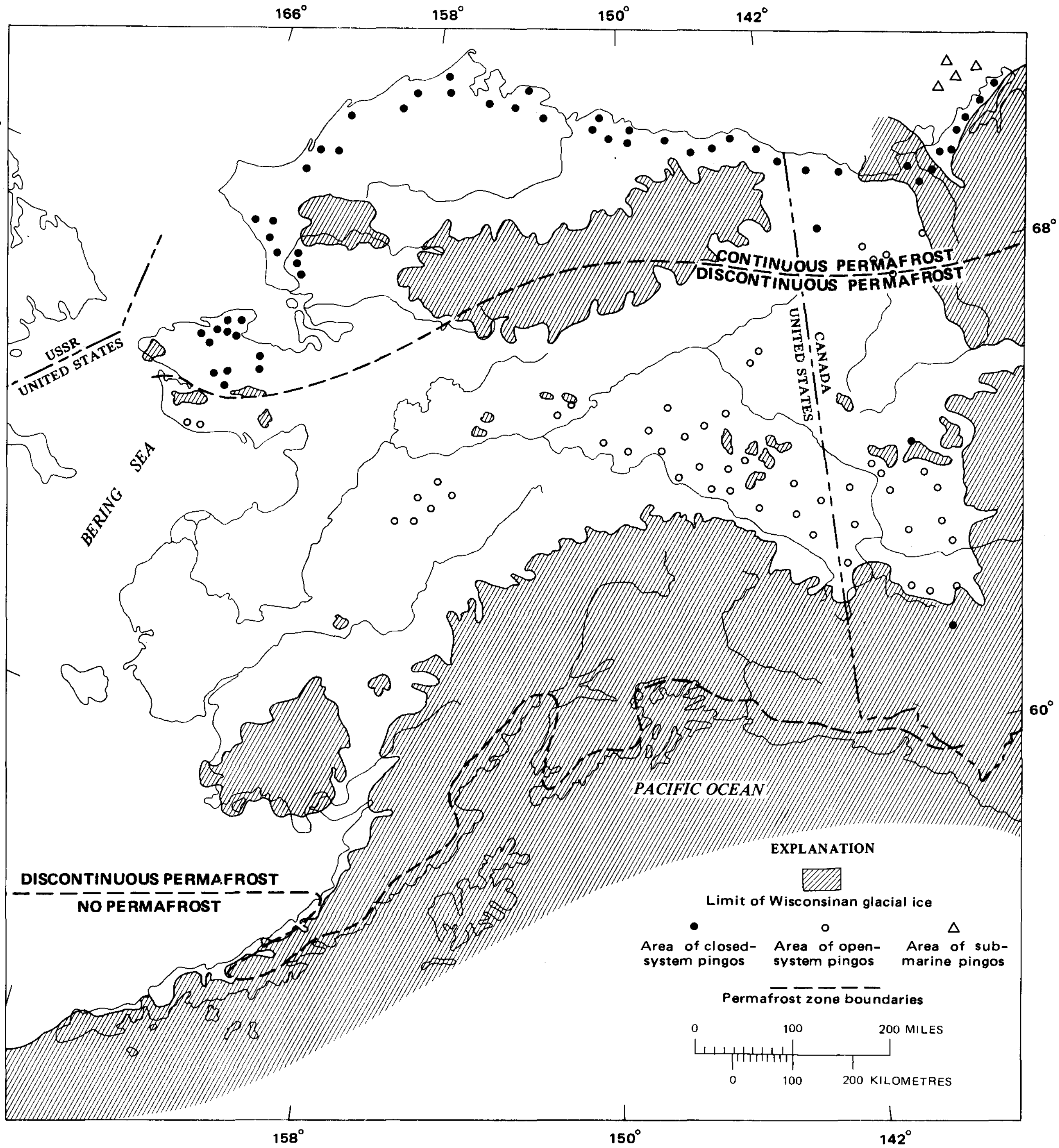


FIGURE 31.—Distribution of open- and closed-system pingos in relation to permafrost zones and areas covered by Wisconsinan glacial ice in North America. Compiled from all published and unpublished sources, including written comments from D. M. Hopkins, J. R. MacKay, I. Tailleur, O. J. Ferrians, Warren Yeend, Helen Foster, and O. L. Hughes in 1972.

1924, p. 8), and although the writer and his colleagues noticed mounds and circular lakes in permafrost of the valleys in central Alaska for many years, the overriding idea that pingos were arctic features probably prevented acceptance or recognition of pingos in forested areas until the late fifties.

It then became obvious that open-system pingos are widespread in central Alaska and are even adjacent to the University of Alaska and the populated environs of Fairbanks (Péwé, 1965a). At least one was even chosen for a house site (Péwé, 1965a, p. 31). Holmes, Hopkins, and Foster were the first to point out the existence, type, distributuion, and abundance of pingos in central Alaska (Holmes and others, 1966, 1968). Subsequently more than 700 open-system pingos have been recognized in central Alaska (Péwé, 1965a; Krinsley, 1965; Foster, 1967) and adjacent Yukon Territory (Vernon and Hughes, 1966; Hughes, 1969; Fyles and others, 1972; Hughes and others, 1972).

From figure 31 it is easily seen that open-system pingos are essentially restricted to discontinuous permafrost zones. Open-system pingos lie near the base of slopes. Evidently water enters the subsurface system from the surface in the nonpermafrost areas upslope. Studies in the last decade indicate overwhelmingly that almost all open-system pingos in northwest Alaska and Yukon lie on the south or southeast-facing slopes of alluvium-filled valleys (Holmes and others, 1968; Hughes, 1969). The reason for this is still unknown, although perhaps there are more opportunities for surface water to enter the ground in the nonpermafrost areas on south-facing slopes.

One of the most interesting features of the distribution of pingos in northwestern North America is the relation of pingos to the glacial border (fig. 31). Almost all open-system pingos lie in the unglaciated part of the country, but a few pingos are known in areas that have been glaciated more than 25,000 years ago. Pingos are extremely rare in areas glaciated within the last 25,000 years; none are known in Alaska and only two are known in Yukon Territory (Hughes, 1969), but perhaps about half of the 1,500 in the Mackenzie delta (MacKay, 1973) are in areas glaciated during the late Wisconsinan. Hughes (1969) speculated that the absence of open-system pingos in the glaciated areas may be attributed to glacial modification of the topography and surficial materials necessary for their origin or to differences in the extent and thickness of permafrost resulting from glacial and postglacial history of the area. The reason for such a distribution of open-system pingos in relation to the glacial border constitutes one of the problems that should be investigated in the near future.

Radiometric dating indicates that most pingos in Alaska are less than 4,000–7,000 years old (Krinsley, 1965). MacKay (1973) showed by careful field measurement over several years that many are probably less than 1,000 years old and some less than 100 years.

One of the remaining problems in need of more research is a study of the water associated with open-system pingos as well as the roll of artesian pressure. In 1968, Lissey began a study of hydrology of open-system pingos near Dawson, Yukon Territory, but no detailed results are yet available (Lissey, 1970). Williams and vanEverdingen (1973) mentioned that one of the problems to be solved is a better understanding of the pressure developed to lift the static load of a pingo, inasmuch as a total force of 6–22 atmospheres is thought to be required. The only work known on the chemical and biological properties of water in a pingo lake is the preliminary study by Lotspeich, Müeller, and Frey (1969) on a lake in east-central Alaska.

GEOCHEMISTRY OF PERMAFROST

A relatively new approach to the study of perennially frozen ground is the chemical investigation of ice and sediments associated with permafrost; this approach appears to hold great potential. The work is being pioneered in Alaska by Jerry Brown and his associates of the U.S. Army Cold Regions Research and Engineering Laboratory (Brown, 1965b, 1966a, b, 1967a, 1969a, b, d; Brown and Johnson, 1965; Brown and others, 1967, 1969; Sellmann, 1967, 1968; Allan, 1969), although the first worker in the field in Alaska was perhaps O'Sullivan (1961, 1966). Brown stated (1969a, p. 1)

> The spatial distribution of cations and anions in near-surface, earthy materials, such as soils and frozen ground, provides a means of interpreting and assessing present and past chemical regimes and geomorphological activities of a given landscape. This technique is particularly applicable to cold, perennially frozen ground in which movement of soluble ions is small and perhaps even insignificant. Therefore, the existing ionic concentration gradients and their lateral dimensions can be indicative of cold regions environments and provide a measure of conditions prior to the formation of the perennially frozen ground.
>
> The distribution of soluble and exchangeable ions in soils, perennially frozen ground, and sediments underlying water bodies is influenced by both the materials and the present and past depositional and leaching environments. For the same material and environment, low concentrations indicate considerable leaching or freshening and high concentrations indicate lack of these active processes or enrichment by ground or surface waters. Silts and clays retain more soluble and exchangeable ions than do sand and gravels. Ionic concentration generally increases with depth, particularly in uplifted marine sediments.

Most of the work in this field has revolved around the distribution of soluble salts of sodium, magnesium, calcium, and potassium in perennially frozen sediments as a reflection of past thermal and leaching regimes. It has been shown that a relatively low concentration of soluble salts in permafrost indicates leaching during an unfrozen state and then refreezing. At Barrow (Brown, 1969a), this low concentration in Holocene marine sediments has been interpreted as freshening due to past lake migration or to local or regional thawing, leaching, and refreezing.

For the Fairbanks area, Sellmann (1967, 1968) indicated an abrupt change in chemical concentrations of extractable cations in permafrost at a stratigraphic unconformity in retransported sediments of Wisconsinan age (fig. 20). Preliminary interpretations suggest thawing and refreezing above the unconformity. Four cores, the longest 25 m deep, of perennially frozen ground from near Fairbanks were examined by Brown (Brown and others, 1967, 1969) and also indicate an increase of soluble salts beneath an unconformity. The soluble salt content is low in the active layer and increases in steplike fashion at depth in the perennially frozen ground.

The writer applied techniques developed by Jerry Brown and P. V. Sellmann to the excellent exposures of frozen ground in the mining cuts near Fairbanks that have long been studied by other means. Perennially frozen samples collected by the writer from critical points in a 40-m vertical exposure on Eva Creek, 16 km west of Fairbanks, were analyzed for soluble salts by Brown and Sellmann at the U.S. Army Cold Regions Research and Engineering Laboratory (written commun., Dec. 28, 1967 and May 19, 1971) by a method described earlier (Brown and others, 1967) and reported in terms of conductivity (μmho). The work is continuing, but preliminary results are presented here, in table 6 and figure 29, to illustrate possible value of such studies and their application elsewhere.

As table 6 shows, the present active layer is characterized by the low conductivity figure of 65–85. It has long been known that the upper 1 or 2 m of permafrost in interior Alaska thawed during early to middle Holocene time and then refroze (Péwé, 1952a, 1968b, 1975). This cycle is supported by the relatively low conductivity

TABLE 6.—*Conductivity and stratigraphy of sediments exposed at Eva Creek gold placer mining cut*

[Conductivity examination made at U.S. Army Cold Regions Research and Engineering Laboratory, Hanover, N.H., by Jerry Brown and P. V. Sellmann]

Thickness (m)	Unit	Conductivity (μmhos) of samples		Age
		1967	1970	
	Surface			
2	Loess		65 85	Holocene
	(dashed lines = present permafrost table)	156	155	
13	(dashed line = past permafrost table) Retransported loess	314 249 252 285 304 317 234 250 410	235	Wisconsinan
1	Silt-forest bed	143		Sangamon
19	Sangamon soil? Loess	100 1,125 1,100	355 965 300 450 300 480 200	Illinoian
1	Silt-forest bed	883		Yarmouth(?)
	Gravel	1,285		Pre-Yarmouth(?)

figure in the perennially frozen Holocene silt, a figure lower than frozen Wisconsinan sediments but higher than current active layer sediments. Leaching was active during the thaw period. Conductivity jumps to 300 just below the Wisconsinan-Holocene unconformity, and all values in Wisconsinan sediments are about that magnitude, supporting the suggestion of an accumulation and freezing period. Thermal "unconformities" within the Wisconsinan sediments could be present but missed in the sampling.

The drop in conductivity from 410 at the base of the Wisconsinan sediments to 143 in the forest layer (Sangamon) is quite striking and supports the suggestion of an interglacial period of no permafrost and perhaps little accumulation. A period when the ground was unfrozen or when little or no loess accumulated would be favorable for leaching (Péwé and Sellmann, 1971).

A sample at the very top of the loess section of Illinoian age yields a conductivity of 100. It is thought that this low figure represents the Sangamon soil and indicates thawing and leaching of Illinoian loess.

The pronounced drop in values in the forest soil layer at the base of the section relative to bounding deposits is similar to the forest layers higher in this section and strongly supports the suggestion of an interglacial or interstadial period with thawing and little deposition.

The figures for conductivity of the sediments near the base of the section are much different from the upper part (table 6). The base of the permafrost in many exposures in central Alaska today, and perhaps in the past, was at the base of the silt deposits, that is at the contact between the overlying loess or retransported loess and the underlying creek gravel. Ground water may have circulated freely in the gravel just under the silt and may have been high in soluble salts. This would account for the high conductivity figures at the base of the silt section, including the forest bed, and the top of the creek gravel. In nearby exposures many ironstone and calcium carbonate concretions occur at the contact between the silt and the gravel. Also, contemporary ground water analyzed from near Fairbanks has a conductivity figure of about 1,000 μmhos. Therefore, perhaps the high conductivity figures at the base of the section at Eva Creek are the result of ground-water modification.

It is perhaps logical to assume, as Brown, Gray, and Webster (1967) mentioned, that the low soluble salt content of fine-grained soil is probably due to removal by ground-water movement (leaching) during a time when the sediments are unfrozen, especially near the surface. Specifically, they found that unfrozen sediments of the present active layer at Fairbanks, which has been leached for many years, have a conductivity value of 25. Also, in periods of rapid, thick accumulation, sediments have less opportunity to be leached than in periods of little or no accumulation. This has been clearly demonstrated in loess in Iowa (Davidson and Handy, 1952, table 5) and Mississippi (Krinitzsky and Turnbull, 1967, p. 9). In areas there where loess accumulates slowly, the carbonate is almost entirely leached out, in contrast to high carbonate content in areas where loess accumulates rapidly.

Preliminary geochemical investigations dealing with ice wedges indicate that the ionic content of the ice is not all from surface meltwater moving down the contraction crack as was earlier assumed but in part is the result of addition of debris from the sides by wedge growth. The percentage of soluble salts increases with depth in the wedges at Barrow. It would be well to continue this work on wedges of different ages, sizes, shapes, and geologic locations elsewhere in Alaska.

Preliminary geochemical analyses of frozen sediments and ice in northern and central Alaska suggest that the chemical investigation of frozen ground in permafrost regions holds a great potential for gross correlations, and perhaps for interpretation and geomorphological and thermal changes in Quaternary time.

PERIGLACIAL PROCESSES AND DEPOSITS

In parts of central, northwestern, and northern Alaska, some of the modification of the landscape is by geologic processes active in a periglacial region. These processes include nivation, frost riving, frost stirring and sorting, solifluction, and choking of small streams with detrital material moved downslope in part by frost action. Both Troll (1948, fig. 1) and Peltier (1950) proposed a specific geomorphic cycle of denudation and wearing down of the landscape by the processes which might be described by Bryan's (1946) term "cryoplanation," although terms such as "altiplanation" (Eakin, 1916, p. 78) and "equiplanation" (Cairnes, 1912a, p. 338–348) suggested by earlier workers in Alaska and the Yukon perhaps are equally appropriate.

Periglacial processes can be grouped into two types: processes such as nivation and frost action that produce erosional landforms and erosional debris, and processes such as solifluction, rubble sheet and block stream, and rock glacier movement that transport this debris down slopes to form depositional landforms. The erosional landforms produced are altiplanation terraces, scarps, cliffs, and tors. The deposits include slope and valley bottom accumulations owing to mass wasting and are termed solifluction deposits, rubble sheets, block fields, and rock glaciers. The term "colluvial" deposits or colluvium of some authors has been used, at least in part, synonymously in Alaska for periglacial deposits (Fernald, 1960, p. 214–219; Hopkins and others, 1960, p. 54; Holmes, 1959b, p. 54; Péwé and Holmes, 1964).

Inherent in the term "periglacial" is that the processes and the deposits under consideration generally occur below snowline. Inasmuch as the position of snowline varies from one part of the State to another, the altitude at which the processes are active also varies. In addition, the position of snowline changed several times during the Quaternary Period. Both present and past variations in snowline are shown along a line extending from Seward Peninsula east to the Canadian border and another line extending from the Gulf of Alaska north to the Arctic Ocean (figs. 8, 9).

In Alaska it may be possible to study the gap that exists between actively forming periglacial deposits and the "fossil" deposits. Therefore, periglacial deposits may occur at various altitudes in Alaska and be of various ages. Variations in location of deposits and activity of the processes occur with changes in latitude and altitude. The writer and his colleagues are analyzing the past and present distribution of periglacial processes and deposits in Alaska.

Altiplanation terraces are the most widespread periglacial erosional form in Alaska (fig. 32); fresh and relict forms occur throughout the central, western, northwestern, and southwestern part of the State. Altiplanation terraces are large bedrock steps or terraces on ridgecrests and hilltops (Eakin, 1916, p. 78); the terraces possess at least one scarp (ascending and (or) descending) and a tread surface. The tread or "flat" area is 10 to several hundred metres wide and long, and slopes from 1° to 15°, parallel to the ridgecrest. Terrace scarps are from 1 to 30 m high. Treads and scarps are cut into all bedrock types, but altiplanation terraces are best developed on closely jointed resistant rocks such as basalt and andesite. They are more poorly developed on granite and tilted sedimentary or low-grade metamorphic rocks. Residual bedrock knobs (tors) project above some terrace treads.

Altiplanation terraces form perhaps a little below the general altitude of snowline. The scarp retreats by nivation, forming a tread, and surficial debris is removed across and over the treads by mass movement. Above snowline, glaciers, cirques, horns, and arêtes form. On isolated ridges and peaks not large enough to support large glaciers, however, altiplanation terraces form above snowline. As terraces form, frost-rived debris is shed as a blanket of mass-wasting material from the terraces down the slopes to creek valley bottoms. Whether this abundant bedrock rubble is transported as solifluction material, rubble sheets, or rubble steps depends largely on the amount of fine material and water.

The widespread altiplanation terraces in Alaska occur at different altitudes and vary in sharpness of form; some are well developed with sharp edges to the

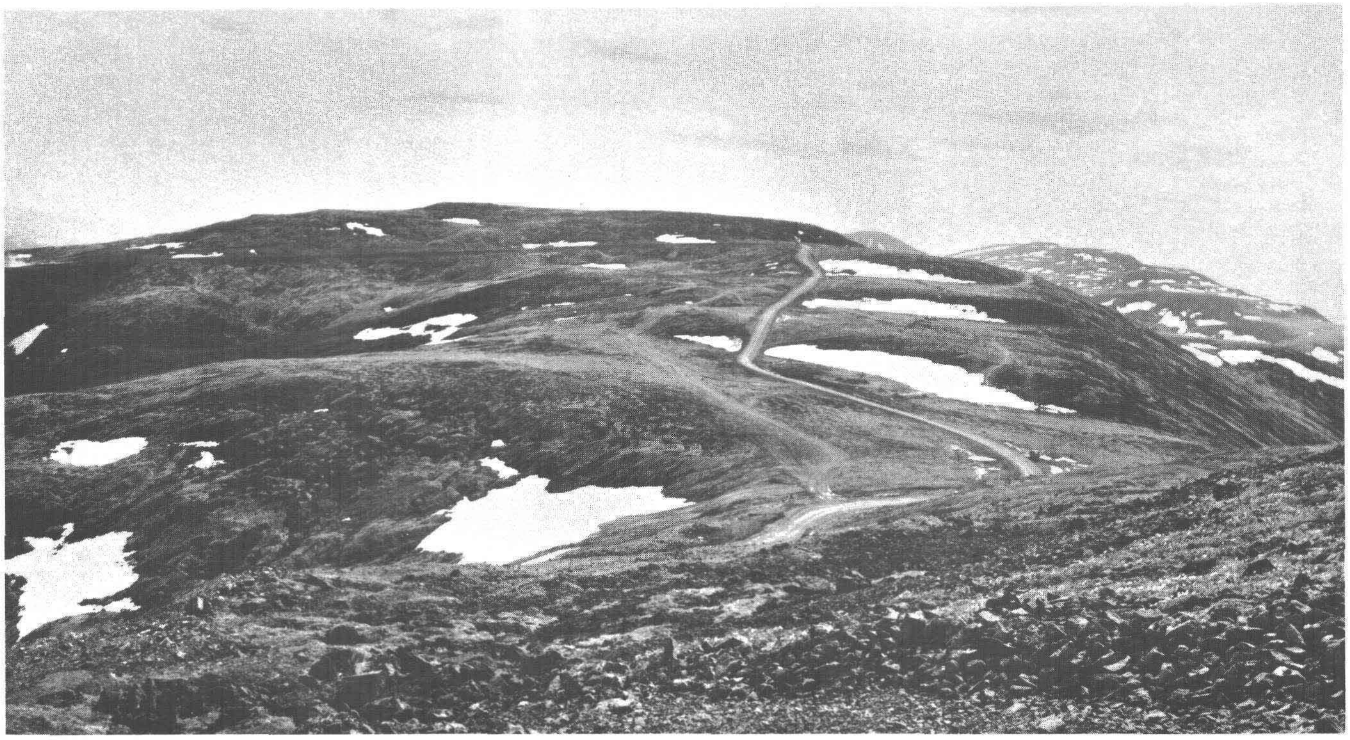


FIGURE 32.—Altiplanation terraces cut into a greenstone ridge of Indian Mountain, Hughes A-2 quadrangle, Alaska. View toward south of transverse snowbanks against north-facing terrace scarps. Photograph No. 2498 by T. L. Péwé, June 25, 1967.

scarps, and some are less "fresh." In all areas the fresher, sharper ones are at higher altitudes, and in some places they can be traced continuously to lower altitudes on the same ridge, where they are older and more rounded.

Well-formed altiplanation terraces occur at 1,100–1,400 m in the Yukon-Tanana Upland. Farther west in the Indian Mountain area near Hughes, they occur at 900–1,000 m above sea level (Eakin, 1916, pl. 7). They are well formed at lower altitudes farther west, for example, at 400 m above sea level near Marshall (Harrington, 1918, pl. V). A plot of the elevations of well-formed altiplanation terraces from east to west in Alaska falls on a line below, but parallel to, past and present snowlines (fig. 8).

It is not known if altiplanation terraces are actively forming in Alaska today. It is known that terraces were formed in Wisconsinan time (Péwé, 1965c, p. 90; Reger and Péwé, unpub. data, 1974), and some may still be slightly active.

One of the problems remaining is the determination of the climatic conditions necessary to produce the terraces. The exact knowledge of such conditions would aid in determining the paleoclimatic conditions under which ancient altiplanation terraces formed. Relict altiplanation terraces have elevations of about 1,600 m in the Yukon-Tanana Upland; at Fairbanks, however, altiplanation terraces of pre-Illinoian age occur at an elevation of 260–900 m (Péwé, 1969b, 1970b).

Most periglacial deposits in Alaska can probably be attributed to the mass-wasting processes described by solifluction. The meaning of the term as introduced by Andersson (1906) is now slightly modified, and Smith's (1949, p. 1499) definition, "progressive downslope movements of presumably saturated detrital material under action of gravity, probably working in conjunction with frost-heave in most instances," is used here. The fact that this material moves in a semifluid condition is suggested by the lobelike or sheetlike flows commonly exhibited on slopes. One of the outstanding features of solifluction is the mass transportation of material over low-angle slopes (Capps, 1919, p. 69; Wahrhaftig, 1951, p. 180; Sigafoos and Hopkins, 1952, p. 183).

Solifluction deposits are widespread in Alaska and consist of a blanket, approximately 0.3–6 m thick, of unstratified or poorly stratified, unsorted, heterogeneous, till-like detrital material of rather local provenance. The debris is either angular or rounded, depending upon the parent material. In actively forming deposits, turf and peat may be incorporated.

In some areas of active solifluction in Alaska, the terrain is characterized by relatively smooth, rounded hills and slopes bearing well to poorly developed solifluction lobes or terraces. If the debris is blocky and angular and fine material is absent, the lobes are poorly developed or absent entirely. Areas in which solifluction lobes are well formed lie almost entirely above or beyond the forest limit (fig. 8). The lobes are at an elevation of 1,200 m in central Alaska on the Canadian border and at progressively lower altitudes to the west, until on the Seward Peninsula they are only a few hundred metres above sea level (fig. 9). Table 7 gives elevations of some active solifluction lobes in Alaska.

In northern Alaska, solifluction deposits are actively forming over a wide area from elevations near sea level to near 1,500 m. Solifluction lobes have been detected on aerial photographs of central, western, and northern Alaska. It should be noted that whereas well-formed lobes are an easily recognizable indication of the process of solifluction, the process may be active and important at altitudes several hundreds of metres lower than where such lobes occur.

Quantitative studies of active mass wasting in Alaska are only beginning (Everett, 1962, 1966). Most studies of solifluction deposits have been made in temperate latitudes where solifluction is no longer active. These studies have proved valuable in the interpretation of past geologic events and paleoclimatic environments. Examination of the widespread stabilized solifluction deposits in Alaska has only just begun, but they hold great promise for a more detailed interpretation of Quaternary events in nonglaciated regions.

At Cape Denbigh (Hopkins and Giddings, 1953, p. 18), along the Firth River in northwest Yukon Territory, in Canada near the Alaska-Canada boundary (Mackay and others, 1961, p. 46), and at the Onion Portage site on the Kobuk River (Giddings, 1967), evidence of solifluction within the past few thousands of years has been helpful in interpreting the geologic history of archaeological sites.

Stabilized solifluction sheets of pre-Illinoian age are widespread in the Yukon-Tanana Upland and are best known in the Fairbanks area. Underlying the extensive loess deposits of central Alaska are one or more sheets of stable solifluction material 0.3–3 m thick blanketing the hillsides (fig. 20) (Péwé, 1965a, fig. 1–2; 1965b, figs. 4–21, 4–22; Blackwell, 1965; Péwé, 1969b). The poorly developed stratification is "dragged out downslope." One of the most thoroughly studied areas of ancient solifluction deposits is on the campus of the University of Alaska. Here a complex series of events is recorded by solifluction layers and ice wedge casts ("fossil" ice wedges) beneath Wisconsinan and Illinoian loess (fig. 33). The two mass-wasting deposits are thought to be early to middle Pleistocene in age, and the ice wedge casts probably are evidence for the oldest permafrost known in Alaska. D. M. Hopkins (oral commun., 1965) observed a similar stratigraphic record in the Tofty mining district 130 km west of Fairbanks. It should be noted that the stabilized solifluction deposits in the Yukon-Tanana area occur at an elevation of 150 m above sea level or even lower, a lowering of at least 900 m from the position of the well-formed lobes present today (table 7). In the absence of tectonic movement, we must assume a dramatic lowering of rigorous

TABLE 7.—*Elevation above sea level of active solifluction lobes in Alaska*

Elevation (m)	Location	Source
	Central and Western Alaska	
1,100–1,200	Yukon-Tanana Upland	Taber (1943, p. 14), Péwé (1952a, fig. 7).
1,400	Central Alaska Range	Péwé (1952a, fig. 8), Wahrhaftig (1958, p. 59).
600–900	Kokrine Hills	Eakin (1916, pl. 7A).
750	Near Hughes, north side of Ray Mountains.	T.L. Péwé (unpub. date, 1970); Eakin (1916, pl. 7A).
200	Nome and vicinity	Sigafoos and Hopkins (1952).
100–150	Cape Denbigh	Hopkins and Giddings (1953, pl. 1).
150	Pribilof Islands	D.M. Hopkins (written commun., 1969).
	Northern Alaska	
Near sea level	Alaska-Canadian border	MacKay, Matthews, and MacNeish (1961, p. 44–46).
500	Alaska-Canadian border	Drew and Shanks (1965).
600–900	North flank of Brooks Range	Holmes (1959b, p. 55).
600	Anaktuvuk Pass	Porter (1966, pl. 18).

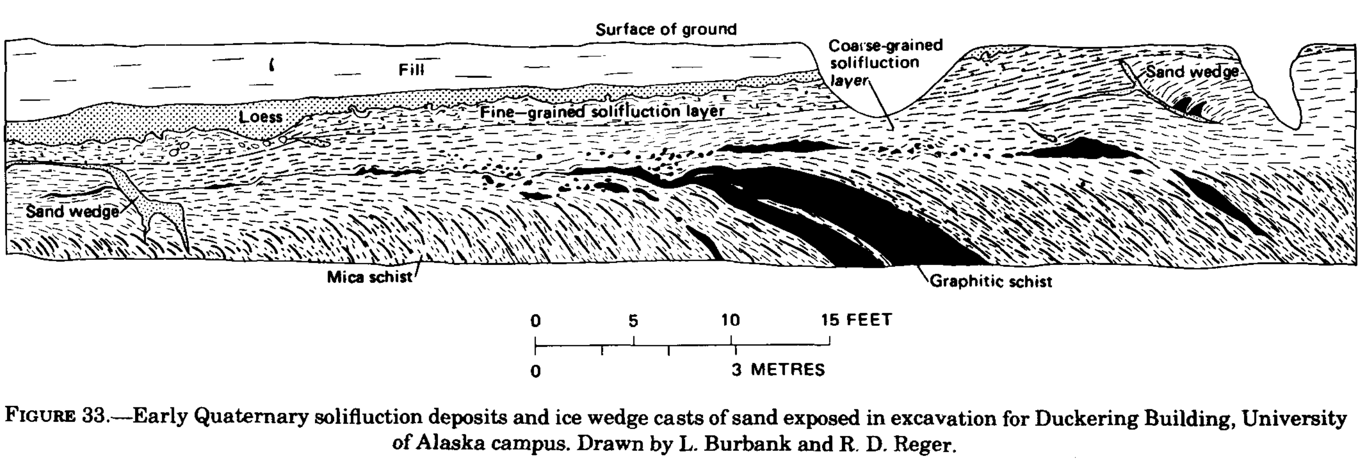


FIGURE 33.—Early Quaternary solifluction deposits and ice wedge casts of sand exposed in excavation for Duckering Building, University of Alaska campus. Drawn by L. Burbank and R. D. Reger.

solifluction conditions on the north side of the Tanana valley.

In many areas the frost-rived debris contains less fine material and less water and consists of angular fragments of well-jointed, resistant rock such as quartzite, basalt, or andesite. Under such circumstances solifluction lobes do not commonly occur; instead, conspicuous sheets or streams of angular rubble form. These rubble sheets consist of angular to semirounded bedrock fragments a few to 60 cm in diameter; they form a blanket 30 cm to 2 m thick on slopes and hills. These sheets may be hundreds of metres long and wide. Active rubble sheets occur in northern and western Alaska, but little is known of the mechanics of movements and distribution.

The best documented inactive rubble sheets in Alaska are on Jumbo Dome on the north side of the central Alaska Range (Wahrhaftig, 1949). Here the evidence indicates at least five periods of rubble-sheet formation on the andesite dome, interspaced with periods of dissection. The sheets are between elevations of 900 and 1,200 m and were active during times of more rigorous climate. The snowline was 370 m lower during Wisconsinan time (Wahrhaftig, 1949, p. 220), and periods of rubble-sheet formation are correlated with glaciations in the Alaska Range. The youngest rubble sheet may be correlative with the late Wisconsinan glacial stage.

In the Amphitheater Mountains on the south side of the central Alaska Range along the Denali Highway, rubble sheets of Wisconsinan age occur at 1,200 m (Péwé, 1965c, fig. 7–39). The rubble was derived from higher slopes during formation of altiplanation terraces during the rigorous climate of the Denali (Wisconsinan) Glaciation (table 2; Péwé, 1961c, D200–D201; 1965c). The rubble overlies till of Illinoian age.

Rock glaciers, tongue-shaped or lobate masses of unsorted, angular frost-rived material with interstitial ice (if active) and with steep lichen-free fronts 10–100 m high, are one of the most spectacular periglacial deposits (fig. 34) but are limited in aerial extent. They are 150–3,200 m wide and 300–1,600 m long, and in Alaska most of them occur in cirques. Rock glaciers are abundant in the Talkeetna Mountains and the Chugach Range and were reported by Capps (1910a) and Moffit and Capps (1911, p. 52) from the Wrangell Mountains (fig. 34). They are widespread in the central Alaska Range (Wahrhaftig and Cox, 1959; Foster and Holmes, 1965).

Active rock glaciers occur below snowline, and a study of topographic maps and aerial photographs indicates that they are at an elevation of approximately 1,600 m on the north side of the Alaska Range, at 1,500 m in the Talkeetna Mountains, at 1,200 m in the Chugach Range, and at 1,150 m in the Wrangell Mountains. Porter (1966, fig. 85) reported active rock glaciers at an elevation of 1,600–1,800 m in the central Brooks Range.

The most detailed study of rock glaciers in Alaska was by Wahrhaftig and Cox (1959) in the central Alaska Range. They defined a rock glacier as a type of glacier formed under the influence of a periglacial climate in an area lacking the net accumulation of snow required for a conventional glacier to form. A permafrost environment must be present (mean annual air temperature colder than 0°C) to enable the snow and water which trickles down into the interstices between the rocks to remain as ice. The abundant supply of coarse blocky debris necessary for formation of rock glaciers is most often found at the base of cirque walls. A periglacial climate with its intense frost action is ideal for frost riving of the fractured bedrock of the cliffs.

Active rock glaciers move slowly. Measurements of one rock glacier in the central Alaska Range by Wahrhaftig and Cox (1959, p. 383) indicate that it moved at the average rate of 73 cm annually from 1949 to 1957.

An active rock glacier is in equilibrium with a climate necessary to produce the frost-rived debris and to permit the interstitial ice to exist. The rock glacier becomes inactive when these climatic conditions are not met and thus loses its steep front, interstitial ice, and forward motion. In Alaska, inactive rock glaciers can be found at lower elevations than the presently active ones, and no doubt they represent the lowering of snowline and changing of other climatic parameters in the past. Wahrhaftig and Cox (1959) carefully showed that an understanding of rock glaciers permits an interpretation of paleoclimatic conditions. They documented many post-Wisconsinan inactive rock glaciers in the Alaska Range. In the central Alaska Range, they stated that the rock glaciers formed in two separate cold periods that occurred after the post-Wisconsinan thermal maximum. Almost all inactive rock glaciers in Alaska occur in Wisconsinan cirques and must therefore be post-Wisconsinan in age; however, at an elevation of 1,200 m along the Denali Highway between the Tangle Lakes and MacLaren River, inactive rock glaciers(?) of Wisconsinan age are reported (Péwé, 1965c, p. 91). In the Anuktuvuk Pass area, Porter (1966, fig. 85) found inactive rock glaciers at elevations from 1,200 to 1,500 m.

Rock glaciers in Alaska are excellent records of late Holocene events, but only in the central Alaska Range have they been used in interpreting the history of the last tens of thousands of years (Wahrhaftig and Cox, 1959). A great opportunity exists in most of the mountains in Alaska to document this span of history carefully by means of rock glaciers.

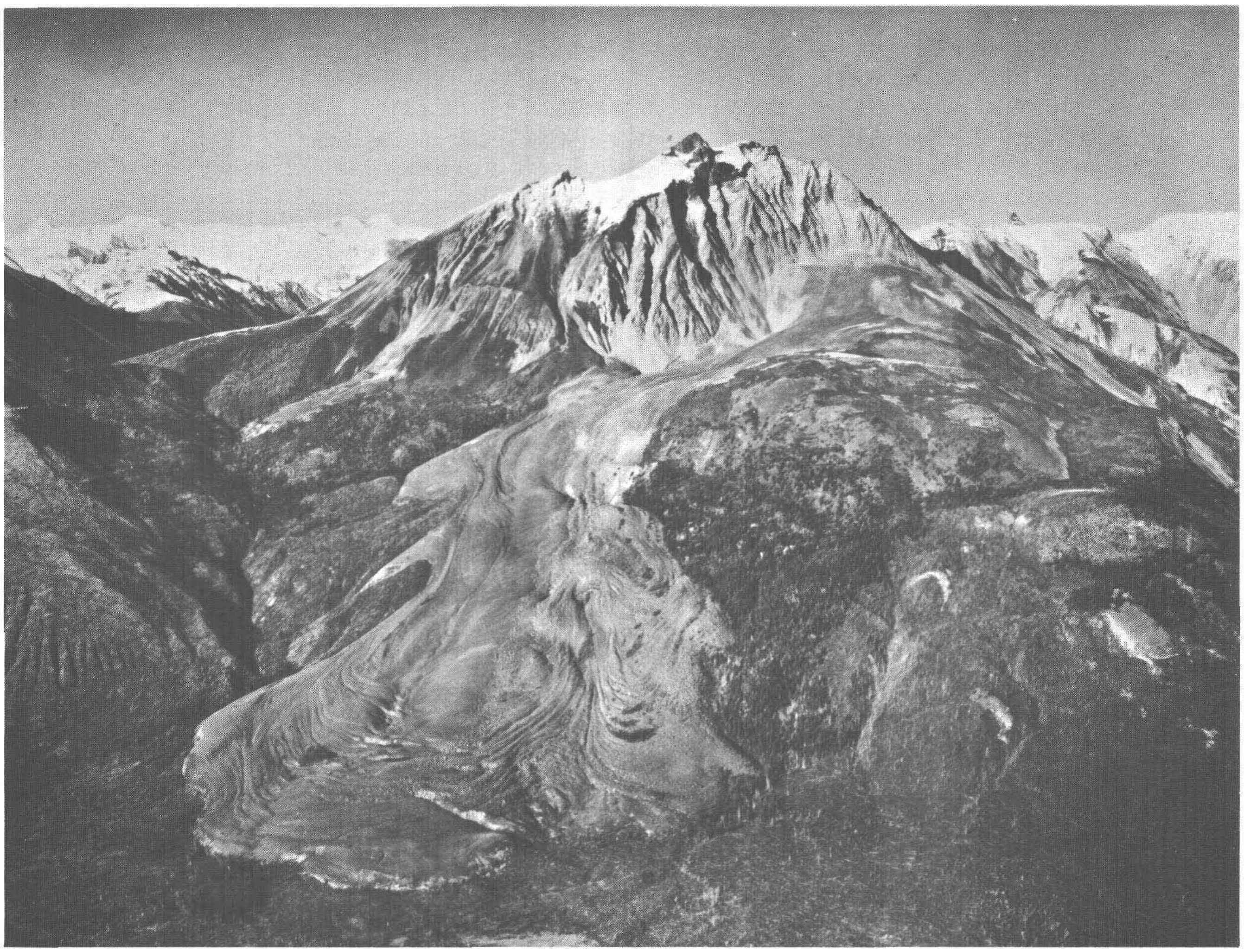


FIGURE 34—Rock glacier on Sourdough Peak in the southern foothills of the Wrangell Mountains 13 km east of McCarthy. View looking north. Photograph by Bradford Washburn.

THERMOKARST FEATURES

One phenomenon restricted to the periglacial landscape is thermokarst topography. The thawing of permafrost creates an uneven topography which consists of mounds, sinkholes, tunnels, caverns, short ravines, lake basins, and circular lowlands caused by melting of ground ice. Thermokarst features differ from other periglacial phenomena inasmuch as they are not formed by the repeated freeze and thaw of bedrock with subsequent breakup of the rock and transfer of the debris; they are the result of thawing of permafrost and associated ground ice, and they form either by warming of the climate or by artificial or natural removal of the overlying vegetation cover.

Thermokarst features have been noted in Alaska for many years. In the Arctic Coastal Plain, modern and ancient lake basins (Black and Barksdale, 1949; Black, 1969b; Britton, 1958a, b; Tedrow, 1969) as well as ravines (Anderson and Hussey, 1963) and polygonal topographic microrelief (Hussey and Michelson, 1966) (fig. 28) are widespread. Sinkholes and lake basins are reported from the Seward Peninsula (Hopkins, 1949) and central Alaska (Wallace, 1948; Péwé, 1965a, 1966c, p. 27). Mounds and pits are abundant in cultivated fields near Fairbanks (Péwé, 1949, 1954).

Despite these observations, no systematic studies of the thermokarst topography have been made on a regional basis such as those done in the U.S.S.R. (See references in Czudek and Demek, 1970.) Several areas in central Alaska appear to exhibit well-developed, widespread thermokarst topography, such as the lake-pitted plains on the south side of the Yukon Flats (pl. 1). The same type of topography in the middle reaches of the Tozitna River may yield basic data on the origin of thermokarst topography and the regional history of permafrost.

Ice wedge casts are one of the features of buried thermokarst topography that have been used in the interpretation of Quaternary climates; however, recognition of other interesting thermokarst features is rare, although ancient slumping and gulleying are known in the Fairbanks area (Péwé, 1952a, 1965a).

GEOBOTANICAL PERIGLACIAL FEATURES

Two periglacial geobotanical features that are well developed in Alaska but have not been studied, even in a reconnaissance manner, are palsas and string bogs, which are geobotanical features related to permafrost and seasonal freezing of the ground; they have been studied in the subarctic in Europe, Asia, and Canada.

Palsas are mounds of peat and ice, 1–7 m high and 10–50 m in diameter, that occur in bogs. They protrude well above the level of the bog, have a rather hard and dry surface, and are generally blown clear of snow in the winter. The segregation of ice in thin layers and in bodies 2 m thick is palsas, plus the accumulation of peat, accounts for their formation. Palsas are thought to belong to the southern region of discontinuous permafrost (Lundquist, 1969). They are well developed and abundant along the Denali Highway west of the MacLaren River valley on the south side of the Alaska Range and are probably widespread in the nearby upper Susitna valley and perhaps in the upper Kuskokwim valley. The grass and peat-covered pingos reported near Bethel in the Yukon-Kuskokwim Delta by Burns (1964) are thought by the writer to be palsas (fig. 31).

Though Alaska is an ideal place to study palsas and to learn more of their origin, age, and relation to the glacial record and permafrost, thus far only one has been investigated, a dissected palsa in the MacLaren River valley. The feature was termed a peat pingo, and a radiocarbon date of 10,565±225 years (GX–0249) (Péwé, 1965c) (table 2) was obtained from the peat at the upper contact with the clear-ice core. This palsa was photographed (Hamelin and Cook, 1967, p. 29) and discussed during the 1965 Alaska-INQUA field excursion. A nearby undissected palsa has also been photographed (Jahn, 1966, p. 93).

String bogs (strangmoore; tourbière réticulée) are boggy areas marked by a gentle undulating surface characterized by long, low serpentine moss ridges alternating with long, water-filled swales. The ridges or "strings" are 1–3 m wide and as much as 1 m high. They trend across the bog generally at right angles to the areal drainage. Troll (1944, p. 72–74) reported them from various parts of Europe, Asia, and Canada.

The only detailed study of their origin was done 40 years ago in Finland by Auer (1920), who believed that they are produced by seasonal ice growth under ridges which are generally blown clear of snow. Different moisture situations permit certain species of moss and other plants to grow on the ridges while other species grow in the wetter swales. Auer concluded that string bogs are an expression of seasonal frost and are not related to permafrost.

There is considerable controversy at present over the origin of string bogs, and other ideas of origin concern their relation to solifluction and to permafrost. Schenk (1966) stated that string bogs are the result of collapse of formerly perennially frozen bogs when permafrost disappears. Rapp and Annersten (1969) strongly objected to this idea.

String bogs are excellently developed and widespread in the lower Susitna River valley and other drainages from the Alaska Range and are well exhibited in bog flats of the western Kenai Peninsula. Drury (1956) studied bogs in the upper Kuskokwim River region and discussed them in detail. However, string bogs are poorly developed, if at all, there and in central Alaska. They are best formed near the southern border of permafrost. Although the writer has examined them from the air and on the ground and although photographs of these bogs were published by Schenk (1966) and in a classic paper on Alaskan bogs (Drury, 1956), no research has been undertaken to determine their origin and relation to the glacial record and permafrost distribution.

SUMMARY

Permafrost is perhaps the most impressive periglacial phenomenon in Alaska. Currently present in about 82 percent of the State, it was more extensive in glacial cold periods and less extensive in past interglacial periods. It is more continuous, thicker, and colder north of the Brooks Range than in central Alaska. Although permafrost is forming today in most of Alaska, much of the deep-lying permafrost is not in thermal equilibrium with the present climate.

Ground ice, the most significant feature in permafrost affecting human activity in the north, is a key factor in the creation of microrelief and macrorelief forms and provides evidence useful in the interpretation of Wisconsinan and Holocene geologic history. Although detailed studies of ground ice have been made, more are needed, and new information is becoming available with increased engineering construction in central and northern Alaska. New information is now available concerning pingos in central Alaska and even under the shelving shores of the Arctic Ocean.

Periglacial processes are active, and periglacial erosional and depositional landforms are widespread in Alaska. The landforms and deposits are mainly Wisconsinan and Holocene in age, but some Illinoian and pre-Illinoian forms and deposits exist and are excellent clues for the reconstruction of geologic history and paleoclimates. Some provide opportunity for more critical work, especially periglacial geobotanical features such as palsas and string bogs.

FLUVIAL DEPOSITS

Some 10 percent of Alaska may be covered by fluvial deposits, if the overlying loess blanket on the terraces is excluded (fig. 1). In addition to low- and high-level terraces and modern flood plains of rivers, huge areas of glacial outwash fans flank most major mountain ranges. Many large tectonic basins (Payne, 1955), such as the valleys of the Kuskokwim and Tanana Rivers, the Yukon Flats, and Yukon-Koyukuk lowland, are filled with 1 to a few hundred metres of fluvial sediments of Quaternary age. Extensive areas of delta deposits occur in western and northern Alaska.

A complex Quaternary history is recorded by the fluvial deposits in Alaska, but only in the last 25 years have systematic investigations of fluvial deposits been undertaken, particularly in connection with various engineering geology projects of the U.S. Geological Survey.

YUKON FLATS

One of the most extensive areas of fluvial deposits in Alaska is in a tectonic basin termed "the Yukon Flats basin" (Miller and others, 1959), which lies along the Yukon River south of the Brooks Range and north of the Yukon-Tanana Upland. According to Williams (1962), the basin includes two physiographic units: the Yukon Flats and the marginal upland. The Yukon Flats is a broad lowland, approximately 23,000 km^2 in extent, which lies along the Yukon River and its tributaries and includes flood plains, terraces, alluvial fans, and small areas of sand dunes. The marginal upland, occupying an area of more than 12,000 km^2 around the edge of the basin, is 30–150 m above the alluvial lowland and consists of a thick sequence of older alluvial deposits of the Yukon and its larger tributaries and of alluvial aprons formed by small streams draining the adjoining highlands. The fluvial deposits of the upland are a few to more than 30 m thick, but in the center of the flats near Fort Yukon, 30 m of fluvial deposits of Quaternary age overlies more than 90 m of lacustrine deposits (Williams, 1960, 1962); the lowest part of the lacustrine deposits contains a pollen flora of late Pliocene or early Pleistocene age (Williams, 1960, 1962).

The high-level alluvium consists of rounded pebble to cobble gravel with lenses of sand and silt. These fluvial deposits are remnants of a system of high river terraces and are late Tertiary and early Quaternary in age (table 3). A very small pollen flora from an exposure of the alluvium at Birch Creek suggests that this high-level gravel is the same age as lake deposits in wells at Fort Yukon (E. B. Leopold, written commun., 1959).

In the central and eastern part of the Yukon Flats, large alluvial fans and related terraces of the Yukon and its tributaries that drain the Brooks Range cover an area of thousands of square kilometres. The higher and older parts of these fans are correlated with the early and middle Pleistocene morainal complexes in the southern Brooks Range (Williams, 1962) (table 3). The stream deposits of the older fans and terraces are, where observed, heavily iron stained and slightly cemented by iron oxide. Younger alluvial fans and related terraces that can be traced to the younger glacial moraines in the Brooks Range are thought to be of middle and late Pleistocene age (table 3). The apices of the younger fans are at lower elevations than those of the older fans, but toward the Yukon River the younger fans appear to bury the older ones.

Major streams today have wide flood plains, and flood plain and low terrace alluvium cover thousands of square kilometres. The alluvium is of late Pleistocene and Holocene age and is still being deposited at the present time.

TANANA RIVER VALLEY

The upper Tanana River valley, upstream from Cathedral Bluffs, is 160 km long and 15–30 km wide. Drainage is poor, and many lakes dot the surface. Fluvial and glaciofluvial sediments have accumulated here for at least the last half of the Quaternary Period; the total thickness of sediments is unknown. Most of the sediments poured into the valley from the Alaska Range on the south in the form of glacial outwash fans and alluvial fans. Wallace (1948) and Black (1951a, 1958) held that much of the sediment in the upper Tanana valley may be lacustrine. The most detailed work in this area was done by Fernald (1965b), who did not recognize widespread lacustrine deposits; he described outwash, alluvium, and sand dunes (table 3).

The middle Tanana Lowland is a 21,000-km^2 basin between the Alaska Range on the south and the Yukon-Tanana Upland on the north and extending from near Big Delta on the east to near Manley Hot Springs on the west. It is a structural basin, the floor of which is below sea level in much of the trough (Péwé, 1958b; Williams and others, 1959; Barnes, 1961; Andreasen and others, 1964). Quaternary deposits 120–180 m thick are largely an accumulation of fluvial and glaciofluvial sediments shed from the rising Alaska Range. This deposition pushed the Tanana River to the north against the Yukon-Tanana Upland (Mertie, 1937, p. 22; Péwé, 1958b). The sediments bury a fairly rugged topography, the hilltops of which now form small knobs above the plain. Well-drained fans 50 km long flank the Alaska Range. The northern part of the basin consists of a swampy flood plain and the terraces of the Tanana and smaller streams. In the western part of the basin, fluvial deposits are covered with a thick blanket of eolian sand.

Péwé, Wahrhaftig, and Weber (1966) suggested that in Illinoian time, when glaciers were more extensive on the north side of the Alaska Range than now, great alluviation occurred in the basin, and several lakes, which still exist today, were formed between the fill and the Yukon-Tanana Upland (Blackwell, 1965). Some of the fill may be related to continuing tectonic activity. In Sangamon time the rivers cut down, forming prominent terraces 30 m high along the Tanana River upstream from Fairbanks and in the huge outwash fan of the Wood River in the southern part of the basin. In Wisconsinan time alluviation was renewed, especially along the glacial streams, and was followed by local dissection in Holocene time.

KUSKOKWIM REGION

In the vicinity of Lake Minchumina, between the middle Tanana Lowland and the upper Kuskokwim region, is a lowland 80 km wide blanketed partly with fluvial sediments. Little is known yet about this region, although the Quaternary history is perhaps similar to that of the upper Kuskokwim region (Fernald, 1960), a 15,000-km^2 lowland and bordering piedmont between the Alaska Range on the southeast and the Kuskokwim Mountains on the northwest (pl. 1). Fernald (1960) showed that more than half of the area is underlain by fluvial deposits of flood plains, outwash plains, and alluvial fans. Fluvial sedimentation on the south side of the basin pushed the Kuskokwim River northward against the edge of the Kuskokwim Mountains. A well at McGrath reveals 70 m of fluvial sediment, and a well at Farewell, near the foot of the mountains, penetrated 110 m of unconsolidated sediments, most of which are fluvial, although a small amount may be glacial (A. T. Fernald, oral commun., April 16, 1969).

The outwash plains and fans are composed of well-rounded boulder, cobble, and pebble gravel that becomes increasingly fine downstream. The fluvial deposits of the Kuskokwim River flood plain are mainly silt and sand with only local beds and lenses of gravel. Considerable organic material is present. Much of the lowland is covered with bogs (Drury, 1956) that are several hundred square kilometres in extent.

The history of fluvial deposits is related to glacial advances from the nearby Alaska Range. In Illinoian time (thought to be Wisconsinan(?) by Fernald, 1960, p. 268) considerable alluviation occurred, and dissection followed in Sangamon time. In Wisconsinan time, more fluvial deposits were laid down, and the major rivers, especially near the Alaska Range, are aggrading today, according to Fernald (1960), p. 269).

KOYUKUK RIVER VALLEY

Extensive lake-dotted lowlands underlain by fluvial deposits occur in the Koyukuk valley. The largest lowland, and the only one that has been investigated, is the Yukon-Koyukuk lowland, where the Koyukuk River joins the Yukon River (Weber and Péwé, 1970) (fig. 35). The Yukon-Koyukuk lowland is an unglaciated area of 10,000 km^2 that was subject to alternating periods of great deposition and erosion. Quaternary fluvial deposits are more than 126 m thick in places, and locally, a few bedrock knobs project above the alluvium. It cannot be ruled out that some of these sediments may be estuarine (Eardley, 1938), because the major rivers in the area are only 30 m above sea level today and much of the bedrock floor is 50–100 m below sea level.

The lowland consists of two major terraces and broad flood plains of the Yukon and Koyukuk Rivers. The terraces, composed of silt and locally of sand, are 10–75 m above the flood plain (Weber and Péwé, 1961, p. D371).

The flood plains of the Yukon and Koyukuk Rivers in this area are classic examples of development of flood plains by meandering rivers under subarctic conditions (Péwé, 1947). The flood plains are composed of four mappable units, each successively older: linear, advanced linear, coalescent, and scalloped; the units are arbitrary phases in the development of the flood plain (fig. 35). The classification of these flood plain units has been extended successfully to the flood plain of the Kuskokwim River (Drury, 1956). The four units can be distinguished on the basis of shape of the lakes, vegetation, shape and position of the unit, depth to permafrost, amount of ground ice and organic material, as well as character of the river bank and distribution of driftwood (Péwé, 1948, p. 8), and can be easily distinguished on aerial photographs and maps.

Precise dating of the events in the area is not yet possible; however, tracing the periods of alluviation to the mountains suggests that deposits forming the high terrace accumulated during Illinoian time (table 3). This was followed by terrace formation with downcutting in Sangamon time. Inasmuch as this area is only 30–60 m above sea level, however, the fluctuations of sea level in Quaternary time may have reversed the period of downcutting from interglacial to glacial time, similar to events in the lower Mississippi River valley (Fisk and MacFarlan, 1955). Also, movement along the Kaltag fault may have complicated the Quaternary history of the valley (Patton and Hoare, 1968). Bones of Pleistocene land mammals have been found in deposits of the oldest terrace, but none have been reported from younger units.

In Wisconsinan time, alluviation occurred and dissec-

FIGURE 35.—High-altitude oblique aerial photograph looking south toward the junction of the Yukon and Koyukuk Rivers. The Koyukuk River (dark color) joins the silt-laden Yukon River (light color) at the right. Flood plain units indicated by number are the (1) linear phase, (2) advanced linear phase, (3) coalescent phase, and (4) scalloped phase. Unit (5) is the high terrace. Photograph by U.S. Army Air Corps, August 24, 1941. From Péwé, (1948) and Weber and Péwé (1961).

tion followed, possibly at the end of Wisconsinan time, forming low terraces. A radiocarbon date of 8,140±300 years (W–472) (Rubin and Alexander, 1958, p. 1479) was obtained from organic material taken from frozen organic-rich silt of the scalloped phase of the flood plain. Therefore, the formation of much of the modern flood plain postdates this time (Péwé, 1962), and the lowest terrace is more than 8,000 years old.

OTHER FLUVIAL BASINS

In addition to the major fluvial basins where geologic investigations have been undertaken, many major and minor river valleys with large areas of fluvial deposits exist (pl. 1) about which little or nothing is known, except for casual references or generalized mapping as reported in the early bulletins of the U.S. Geological Survey devoted mainly to bedrock geology. Little is known, for example, of the fluvial deposits in the Innoko Lowland of western Alaska or of the Holitna lowland in the middle Kuskokwim region. The Kobuk and Noatak River valleys in the western Brooks Range contain extensive fluvial sediments, especially in their lower reaches, but to date mostly reconnaissance information (Fernald, 1964) is available.

NORTHERN ALASKA

The Arctic Slope of Alaska contains extensive fluvial deposits about which little is known except for recent studies along the route of the Alaskan oil pipeline. The two physiographic provinces that lie north of the Brooks Range and compose the North Slope are the Arctic Foothills and Arctic Coastal Plain provinces (pl. 1).

The Arctic Coastal Plain province is almost completely blanketed by the Gubik Formation of Pleistocene age (Gryc and others, 1951, p. 167; Black, 1964). Only part of the Gubik Formation consists of fluvial sediments (Coulter and others 1960; Faas, 1966); most of it is composed of unconsolidated marine deposits.

Fluvial deposits of Pleistocene age in the Arctic Foothills province consist of numerous terrace deposits along the major streams. In the Utukok-Corwin area of northwestern Alaska, Chapman and Sable (1960, p. 128–129) reported Quaternary gravel deposits 4.5–7.5 m thick, capping terraces 6–150 m above major rivers. As many as three levels of terraces are recognized. The position of the terraces may indicate uplift or changes in the fluvial regimen caused by glaciation during Quaternary time.

Conspicuous terraces covered with gravel exist along the middle and upper Colville River. In the eastern part of the Arctic Foothills province, the major rivers also are flanked by Pleistocene terrace deposits, but no detailed studies have been made.

DELTAS OF YUKON, KUSKOKWIM, AND COLVILLE RIVERS

The delta of the Yukon-Kuskokwim Rivers is a large area of fluvial sedimentation that undoubtedly records a complex history of alternating marine and freshwater deposition, probably like the Mississippi River Delta.

The delta of the Yukon-Kuskokwim Rivers, about 62,000 km^2 in extent, is a vast lowland 15–60 m above sea level and is termed the "Bethel Basin" by Miller, Payne, and Gryc (1959, pl. 1). Local bedrock knobs protrude through the sediments; some are of the basement rock, and others are volcanic cones of Quaternary age (Coonrad, 1957; Hoare and Condon, 1971).

Subsurface data are needed to interpret the history of these deltaic sediments; to date, only 14 water wells, one attempted oil well, and two core tests have penetrated the delta in the Yukon-Kuskokwim area. The 14 water wells near Bethel (Waller, 1957) penetrated as much as 184 m (Feulner and Schupp, 1964) of silt and fine sand, with sparse gravel layers, without reaching bedrock. The presence of wood chips and bark in the sediments suggests freshwater or estuarine deposits. A wood fragment, believed to have come from just below the base of permafrost, has been dated as older than 34,000 years (W–1287) (Feulner and Schupp, 1964). The Pan American Oil Co. drilled a well near Napatuk Creek and penetrated 296 m of Quaternary(?) sediments before reaching old rocks (William Van Alen, written commun., Mar. 14, 1963). Tests of the Pleistocene foraminifer, *Elphidium orogonese,* were found at a depth of 195 m in this well. The Quaternary sediments, which contained wood chips and "snail" shells, have not yet been studied in detail, but no doubt many of the sediments are of marine origin, deposited during marine transgressions.

The subsurface of the delta of the Colville River in northern Alaska is even less known. It is considerably smaller than the Yukon-Kuskokwim Delta and undoubtedly frozen to greater depths. Walker and his associates are currently studying the surface of the delta and its erosional processes (Walker and Morgan, 1961, 1964; Walker and Arnborg, 1966; Walker, 1967; Walker and McCloy, 1969).

AURIFEROUS GRAVEL OF CENTRAL ALASKA

Perhaps the most famous of the fluvial deposits of Alaska and adjoining Yukon Territory, Canada, are the late Tertiary(?) to middle Quaternary auriferous creek and river gravels now buried by frozen silt or other sediments. These gravel deposits are almost entirely confined to unglaciated central Alaska and, in most instances, are buried by retransported loess. Although a large part of the geologic literature on central Alaska is associated in some way with placer gold deposits, emphasis has not been on the Quaternary history of the deposits. Since 1945, the emphasis has shifted.

Not all is yet published, but considerable material has been assembled on Quaternary stratigraphy of nonglacial fluvial deposits in the Yukon-Tanana Upland (Péwé, 1952a, 1958b, 1965a, 1975; D. M. Hopkins, 1963, unpub. data), the Seward Peninsula (Hopkins, 1963; unpub. data), the Platinum area (Hopkins, field notes), and the adjoining Klondike area near Dawson, Yukon Territory (O. L. Hughes, written commun., 1962).

The lithology of the buried gravel varies widely depending upon the locality. Cobb and Kachadoorian (1961), among others, described the auriferous gravel deposits. The larger streams typically have rounded river gravels, but the hundreds of small creeks have gravel that is characteristically angular and poorly sorted. Most of the gravel represents solifluction material that has been transported only short distances by stream action. The placer gold generally lies on bedrock or in the basal gravel, but in some places gold may be concentrated in the gravel section (Péwé 1952a). In glaciated areas on or in stratified glacial drift (Mertie, 1925, p. 241), the gold may be on "false" bedrock. The thickness of gravel varies from several centimetres to at least 74 m, the latter recorded under 57 m of silt near Fairbanks (Péwé, 1958b, boring log No. 78).

The Fairbanks area is one locality where the auriferous gravels have been studied (Péwé, 1952a, 1958b, 1965a, 1975; Anderson and Johnson, 1970). In late Tertiary or early Quaternary time, creeks in the area began to alluviate, and rich gold placers were buried under 3–75 m of coarse, very angular, local, sandy gravel. This gravel has been stained brown by percolating ground water. The alluviation may have been caused in part by tectonic action and in part by raising of local base level, the Tanana River, as a result of glacial advances in the nearby Alaska Range. This early period of gravel deposition was followed by erosion and removal of most of the gravel. Streams reconcentrated much of the gold in the earlier placers and deposited additional gold placers. Most of the stream channels of the second gold placer concentration are offset from the location of the earlier channels, and therefore some of the first placer accumulations still exist as fragmentary bench deposits (Péwé, 1965a, fig. 109) (fig. 20). A second cycle of gravel alluviation followed. The younger gravel is not stained as dark a brown as the older deposit and so is easily differentiated from the older in gold dredge tailing piles of the Fairbanks area. Both gravel deposits are unconformably overlain by loess, or if in valley bottoms, by retransported loess rich in organic remains (fig. 20).

The younger gravel deposit appears to be definitely Quaternary in age because it contains the tusks and bones of mammoths. Identifiable wood remains are rare, but white spruce is recorded. The poorly sorted, angular gravel grades into, and in some places is overlain by, solifluction deposits thought to have originated under periglacial climatic conditions.

In almost all creeks, both gravel deposits lie on a 3-cm- to 2-m-thick auriferous clay layer thought to result from decomposition of the underlying bedrock. Whether the clay resulted from pre-Pleistocene weathering or formed by percolation of ground water during Pleistocene time is currently being studied by the writer and his associates. (Bell, 1974)

If the age of the gravel is Quaternary, it is early to middle Quaternary because the deposits are deeply buried under two or more reworked silt deposits of considerable antiquity. Because of the antiquity, faunal content, and probable origin under rigorous climatic conditions (solifluction), the older gravel deposit of the creeks in the Fairbanks area is thought to be Pliocene to Pleistocene in age, and the younger gravel deposit is thought to be early to middle Pleistocene (Péwé, 1975).

In some glaciated areas, buried placers and associated gravel deposits are known to be Quaternary in age, although details of their ages are not yet available (Mertie, 1925, p. 241). O. L. Hughes (written commun., 1962) of the Geological Survey of Canada stated that the high-level auriferous gravel deposits of the Klondike at Dawson, Canada, near the Alaska-Canada border are pre-Pleistocene in age, but the auriferous gravel deposits of the inner valleys of the area are early Pleistocene.

LACUSTRINE DEPOSITS

Known lacustrine deposits are relatively limited in Alaska, and only one large area, the Copper River Basin, exhibits widespread, well-developed lake deposits. In the early days of geologic reconnaissance in Alaska, several workers reported Quaternary lake deposits throughout the central part of the State. This concept was based on the presence of a widespread blanket of well-sorted tan silt on the valley slopes, valley bottoms, and on the tops of low hills throughout central Alaska. Spurr (1898, p. 200–230) was the first to consider some of the silt in the Yukon River area as lacustrine in origin. Prindle (1913, p. 50) extended this concept to the upland silts of central Alaska.

CENTRAL ALASKA

The strongest support for the existence of widespread lake deposits in central Alaska came from Eakin (1916, p. 73; 1918, p. 45), who postulated two vast inland lakes. One lake is thought to have existed in the Yukon Flats and Porcupine River valley. It was supposed to have been formed when the middle Yukon, which was assumed to have flowed north up the Porcupine to the Arctic, was blocked by glaciers in the present upper regions of the Porcupine River. This hypothetical lake later drained to the southwest by cutting through a bedrock constriction near Rampart. The other lake was created when the Tanana-Nowitna drainage, supposed to have flowed south to the Kuskokwim River basin, was blocked by glacier tongues extending northwest from the Alaska Range. Later this lake drained by cutting through a bedrock ridge near Ruby.

The existence of these large lakes was seriously questioned by A. H. Brooks (in Eakin, 1916, p. 9). The concept of lacustrine origin for the upland silt in central Alaska is considered invalid for several reasons: The glacier barriers cannot be demonstrated to have existed; shorelines and deltas are absent; the silt is present above the 360-m level, the upper level of the lakes; no stratification exists in the silt; little clay is present; no lacustrine fossils are found in the upland silt; and mud cracks and ripple marks are unknown (Péwé, 1955, p. 715–716).

Although the middle and late Quaternary deposits, now known to be loess, around the Yukon Flats do not indicate a lake at that time, it is noteworthy that Williams (1960) reported lake sediments 45 m below late Quaternary deposits in a borehole at Fort Yukon. Williams believed that the presence of at least 89 m of relatively uniform, fine-textured sediments and the topographic form of the basin suggested a large lake; palynology indicates a late Tertiary or early Quaternary age. (See p. 67.)

Instead of a lake formed in the Yukon Flats and Porcupine River valley in Wisconsinan time by glacier blockage in the upper regions of the Porcupine, recent work indicates that a different situation existed. In Wisconsinan time, glaciers in the Yukon Territory blocked river drainage in the upper Porcupine, and a large lake formed in Old Crow Flats near the Alaska border (O. W. Geist, 1949, T. D. Hamilton, 1970, O. L. Hughes, 1971, all oral communs.). The lake overflowed a bedrock divide and drained largely to the southwest down the Porcupine River into the Yukon Flats.

Black (1951a, p. 100; 1958,p. 79) stated that a lake about 4,700 km^2 in extent existed in the upper Tanana valley from Tetlin to the Canadian border and drained southeast into Canada. The silt and sand deposits in the area were therefore assumed to be lacustrine in origin. Fernald (1965a, b), however, believed that the sand and silt, as well as the gravels of the upper Tanana valley area, are mostly glaciofluvial sediments and represent aggradation of the valley in middle and late Quaternary time rather than deposition into an extensive lake.

NENANA RIVER VALLEY

A small deposit of lacustrine clay in the Nenana River Gorge in the central Alaska Range is well documented (Wahrhaftig, 1958, p. 34–36). During the retreat of ice of the Healy Glaciation (table 2), the Nenana River gorge was occupied by a lake 0.5 km wide and 15 km long; this has been termed Lake Moody. The lake deposits are as much as 75 m thick and consist of blue and yellowish-gray, varved silty clay. Delta deposits of coarse sand and gravel interfinger with the clay. The Alaska Railroad traverses the clay for several kilometres, and thawing of this perennially frozen clay produces serious landslide conditions (Wahrhaftig and Black, 1958, p. 77–109).

COPPER RIVER BASIN

Lake sediments are widely distributed in the Copper River Basin, an intermontane basin about 14,000 km^2 in extent, bounded by the Chugach, Wrangell, and Talkeetna Mountains and the Alaska Range (fig. 36). It is termed a basin in this report, although listed as the Copper River Lowland by Wahrhaftig (1965). The basin is drained by rivers which flow in canyons (pl. 1) through the surrounding mountains that support large glaciers today and that contained much larger glaciers in the past. During Pleistocene glacial cycles, glacial ice repeatedly blocked the exits from the basin, producing huge lakes (fig. 36).

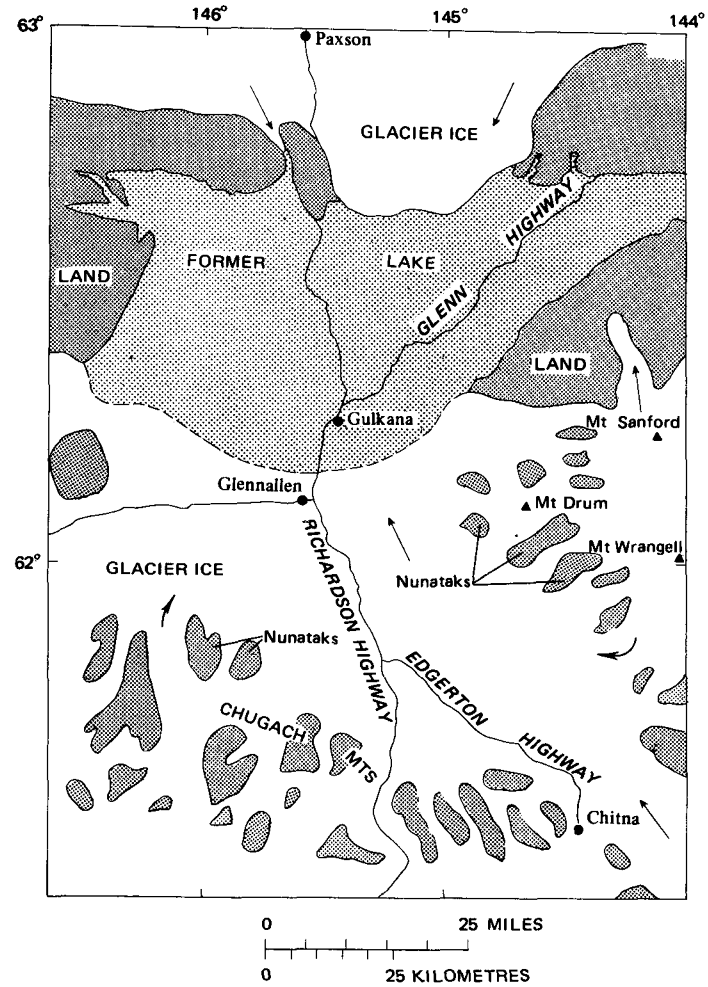


FIGURE 36.—Extent of a lake in the Copper River Basin during the last major Wisconsinan glaciation. Modified slightly from Coulter, Hopkins, Karlstrom, Péwé, Wahrhaftig, and Williams (1965).

The lacustrine sediments in the Copper River Basin were first reported by Mendenhall (1900, 1905) and by Schrader and Spencer (1901, p. 74), but detailed studies of the Quaternary glacial, volcanic, and lacustrine deposits were not undertaken until the 1950's. Only preliminary accounts of these studies were published (Ferrians and others, 1958; Nichols, 1956, p. 11–12; 1961, 1963; Ferrians, 1963a, b; Ferrians and Schmoll, 1957; Ferrians and Nichols, 1965; Nichols and Yehle, 1961a, b; 1969).

The Copper River and its tributaries have dissected the Quaternary sediments of the basin in Holocene times and now flow in steep-walled valleys and canyons as deep as 170 m. Lacustrine, glacial, fluvial, and volcanic mudflow deposits are well exposed in the valley walls (Ferrians and others, 1958). A generalized stratigraphic section of the deposits in the northeastern part of the Copper River Basin is shown in figure 37 (O. J. Ferrians, unpub. data, 1964).

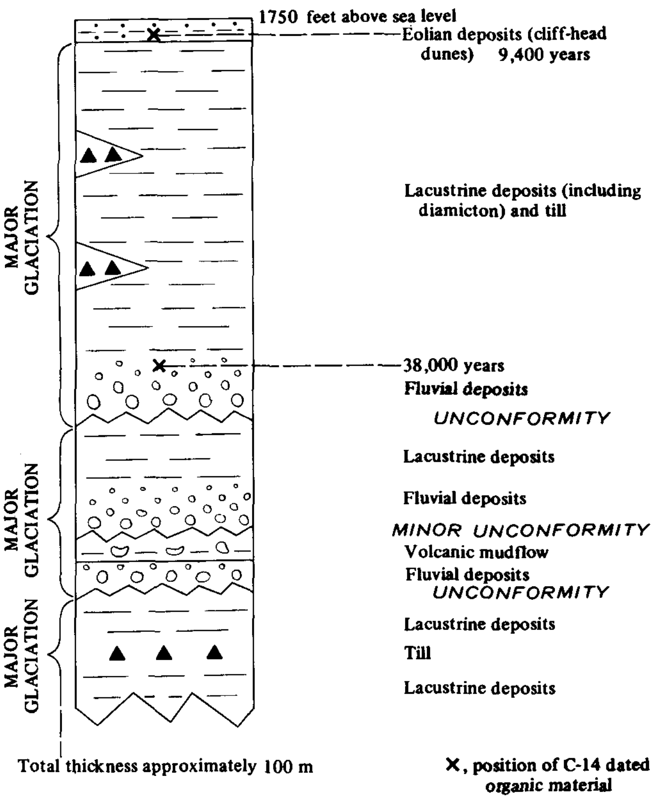


FIGURE 37.—Generalized stratigraphic section, northeastern part of Copper River Basin. Compiled by O. J. Ferrians.

The lacustrine deposits are, in many places, finely laminated, rhythmically bedded sand, silt, and clay with a few layers of volcanic ash (Nichols, 1960). The laminated silt and clay deposits are gray to bluish gray and dry with a blocky fracture. In some places the beds are folded by subaqueous sliding (Nichols, 1960). Many of the fine-grained deposits contain ice-rafted pebbles, cobbles, and boulders, and these grade laterally into well-sorted sand and gravel. Some of the beds were described as diamiction deposits (Ferrians, 1963a, b). Ripple marks are common in the ancient lake sands. The lake sediments deposited during the last major glaciation are as thick as 45 m in the lower parts of the basin, but they are very thin along the margin of the basin. More than 5,000 km^2 of the basin floor is underlain by these deposits.

Apparently, a large lake formed in the basin during each major glaciation that blocked the rivers. During some pre-Wisconsinan glaciations, however, the ice expanded until it filled the basin, overriding previously deposited lake sediments and incorporating them into glacial material. Ferrians (oral. commun., 1962) found evidence in the northeastern part of the Copper River Basin of three major periods of deposition of lake sediments. No basis exists at the present time for relating these major periods of lake sedimentation to glacial sequences elsewhere in Alaska (table 2), except that the most recent episode began more than 38,000 years ago and ended shortly before 9,400 years ago (Ferrians and Schmoll, 1957). Ferrians believed that the three stratigraphic units of lake sediments probably represent episodes of late and early Wisconsinan and Illinoian age (written commun., 1962), whereas the author views them as representing Wisconsinan, Illinoian, and pre-Illinoian episodes.

In the southeastern part of the Copper River Basin, Nichols (oral commun., 1965) found evidence of three periods of lake-sediment deposition that correspond to those of Ferrians and, in addition, two or more older periods. The record of the last lake is clearest, not only because it is more recent, but also because the basin was not completely filled with glacial ice during this interval, and shorelines and deltas remain. The highest shorelines of this age reported by Ferrians and Schmoll (1957) in the northeastern part of the basin lie at an altitude of 750 m, but D. M. Hopkins and D. R. Nichols (unpub. data, 1954) observed lake sediments and shoreline features as much as about 975 m high in the western and northwestern parts of the basin. A radiocarbon date gives a minimum age of 9,400 years for the former lake at Gakona in the northeastern part of the Copper River Basin (Ferrians, 1963a, p. 120).

The glaciolacustrine deposits are now perennially frozen and presumably became so as the lake withdrew (Nichols and Yehle, 1961a, p. 1080–1081; Nichols, 1966). These lake beds are rich in ice and cause serious construction and ground-water problems (Nichols, 1956, p.11–12). The ice-rich lake clays would present one of the most serious problems in the construction of an oil pipeline from northern Alaska to Valdez (Lachenbruch, 1970b).

COOK INLET

Lake clay has been reported from the Cook Inlet area. Near Anchorage, the Bootlegger Cove Clay, a light-gray to dark-greenish-gray silty to pebbly clay ranging in thickness from 1 to 75 m (Miller and Dobrovolny, 1959, p. 41), underlies much of the local sand and clay (Trainer and Waller, 1965) and was the cause of the destructive slides during the 1964 earthquake (Hansen, 1965; Shannon and Wilson, Inc., 1964). Karlstrom (1964) and Schmidt (1963) believed that the clay was lacustrine or estuarine in origin, but detailed studies of its Foraminifera obtained from continous drill cores by P. J. Smith (Hansen, 1965, p. A20) indicate that the entire unit is of marine origin. Schmoll, Szabo, Rubin and Dobrovolny (1972) demonstrated that the clay is 14,000 years old and not middle Wisconsinan age as previously thought (p. 24).

According to Karlstrom (1960b, fig. 154.1), a glacial lake 13,000 km^2 in extent occupied the upper Cook Inlet and lower Susitna valley in late Wisconsinan (Naptowne) time and deposits from this lake, "Lake Cook," covered the lowland and moraines of much of the Kenai lowland and the lower part of the Susitna valley (Karlstrom, 1964, 1965, fig. 9–47; Karlstrom and others, 1964). However, the study of drill cores of the Bootlegger Cove Clay discussed refutes this interpretation.

MIDDLE KOYUKUK RIVER VALLEY

Large lakes are known to have existed in the Middle Koyukuk River valley near Allakaket in late Pleistocene time (pl. 1). Lake clays and drainage anomalies were noted by early workers. Hamilton (1969) outlined parts of these ancient lakes, among them two large ones, 780 and 520 km² in area, which occupied part of the middle Koyukuk valley in nearly Wisconsinan time. These lakes were ponded by glaciers pushing south from the Brooks Range in Illinoian and Wisconsinan time.

Much work remains to be done on the drainage anomalies and the determination of the extent of former lakes. R. D. Reger and T. L. Péwé (unpub. data, 1969) mentioned lake clay associated with Illinoian moraines along the northeastern base of Indian Mountain, 67 km southwest of Allakaket. Modern topographic maps, aerial photographs, and the use of helicopters will permit efficient study of these swampy lowlands. These studies will augment knowledge of the Pleistocene events in this area and will aid in planning roads and other construction on the lake clays, which are extremely susceptible to frost action.

WESTERN BROOKS RANGE

A large Wisconsinan proglacial lake, 60–70 km across, is thought to have formed in the Noatak River valley and its tributaries, between the De Long and Endicott Mountains of the western Brooks Range (Coulter and others, 1965). Most information concerning the lake comes from aerial photograph interpretation, and a great need exists for detailed information on character, extent, origin, and age of this lake, or lakes.

SUMMARY

Pleistocene lakes have been postulated for many areas of Alaska. Detailed work, however, indicates that the sediments of supposed lacustrine origin in central Alaska and the Cook Inlet area can best be ascribed to other origins. Lake deposits are present in the Copper River Basin and present a fine record of late Pleistocene time, but most of the work on these sediments is as yet unpublished. Widespread lake deposits may have formed in the western Brooks Range area when glacial advances blocked drainages. The history and extent of these deposits have not yet been studied.

MARINE DEPOSITS

Alaskan marine deposits of Quaternary age now above sea level are limited to a narrow strip of land along the present coast, except in northern Alaska and perhaps in the Yukon-Kuskokwim Delta area. Most of the deposits were laid down during episodes of higher stands of the sea.

Quaternary marine deposits were recognized in Alaska at the turn of the century (Spurr, 1900; Schrader, 1904), especially on the North Slope (Meek, 1923), in southwestern Alaska, and on the Seward Peninsula (Dall, 1920). However, these isolated reports and later faunal studies were not, for the most part, placed in a firm stratigraphic perspective until the late 1950's and 1960's (Hopkins, 1959b, 1965, 1967a, b; Hopkins and MacNeil, 1960; Hopkins and others, 1960; Hopkins and others, 1962).

Hopkins (1967a) summarized existing knowledge of the distribution, correlation, paleontology, and paleoclimatic and tectonic implications of the Quaternary marine deposits of Alaska, and the following short summary is based mainly on his work. He divided the Quaternary marine deposits of Alaska into seven distinct ages on the basis of flora and faunal content, stratigraphic and geographic position, and paleomagnetic information. Deposits of each age represent a higher eustatic sea level and, therefore, marine sequences of transgression and regression. The marine transgressions affected wide areas simultaneously and have been given provincial names. He then attempted to correlate these transgressions and glacial sequences in western Europe and central United States, although only for the later part of the record because the ages of pre-Wisconsinan events are not precisely known for the classical areas (table 8).

BERINGIAN TRANSGRESSION

The type locality of the Beringian transgression is near Nome (fig. 38). The deposits consist of a few metres of gold-bearing and richly fossiliferous sand and clay (Submarine Beach), of marine origin, covered by till of the Iron Creek Glaciation (table 2) and are characterized by a molluscan fauna that has a decidedly modern aspect (although some assemblages are unlike any modern ones). Other bodies of marine sediments tentatively assigned to the Beringian transgression on the basis of their marine fauna are the lower part of the Gubik Formation (Schrader, 1904) on the lower Colville River and possibly at Skull Cliff near Point Barrow on the Arctic Coastal Plain; deposits of clay and sand 6–9 m above present sea level near the mouth of the Kivalina River; deposits on the Pribilof Islands (Cox and others, 1966) and on some of the Aleutian Islands; and some of the upper part of the Yakataga Formation near Lituya Bay and on Middleton Island. Hopkins (1967a) sug-

TABLE 8.—*Quaternary marine transgressions recorded on Alaskan coasts*
[After Hopkins (1967a)]

Transgression	Type locality	Altitude of shoreline	Climate as compared with the present	Archaeological or radiometric dating	Correlation: North America	Correlation: Europe
Krusensternian	Recent beach ridges at Cape Krusenstern.	Within 2 m of present sea level for deposits <4,000 yr old.	Same	<5,000 yr at Cape Krusenstern; up to 10,000 yr for terraces along Gulf of Alaska coast.	Late Wisconsin and Recent.	Late Würm and Recent.
Woronzofian	Bootlegger Cove Clay near Point Woronzof, Anchorage area (Miller and Dobrovolny, 1959). (See p. 77.)	Probably a few metres below present sea level.	Water colder; air colder.	<48,000 yr; >25,000 yr.	Middle Wisconsin Interstade.	Middle Würm Interstade.
Pelukian	Second Beach at Nome (Hopkins and others, 1960).	Two distinct high sea-level stands at +7-10 m.	Water warmer; air slightly warmer.	Ca. 100,000 yr.	Sangamon Interglaciation.	Broerup Interstade(?) and Riss-Würm Interglaciation.
Kotzebuan	Marine beds below Illinoian drift along eastern shore of Kotzebue Sound (McCulloch and others, 1965).	Probably ca +20 m.	Water same; air unknown.	170,000 yr; 175,000 yr.	Pre-Illinoian Interglaciation.	Mindel-Riss Interglaciation.
Einahnuhtan	Einahnuhto Bluffs, St. Paul Island (Cox and others, 1966).	Probably ca. +20 m.	Water same; air unknown.	<300,000 yr; >100,000 yr.		Pre-Mindel Interglaciation.
Anvilian	Third Beach–Intermediate Beach at Nome (Hopkins and others, 1960).	Probably much higher than Kotzebuan and Einahnuhtan: <+100 m; >+20 m.	Water warmer; air warmer.	Probably <1,900,000; >700,000.	Middle Pleistocene Interglaciation.	
Beringian	Submarine Beach at Nome (Hopkins and others, 1960).	Two distinct episodes during which sea level was higher than at present but probably lower than Anvilian sea level.	Water much warmer; air much warmer.	Last episode ca. 2,200,000 yr on St. George Island.	Late Pliocene.	

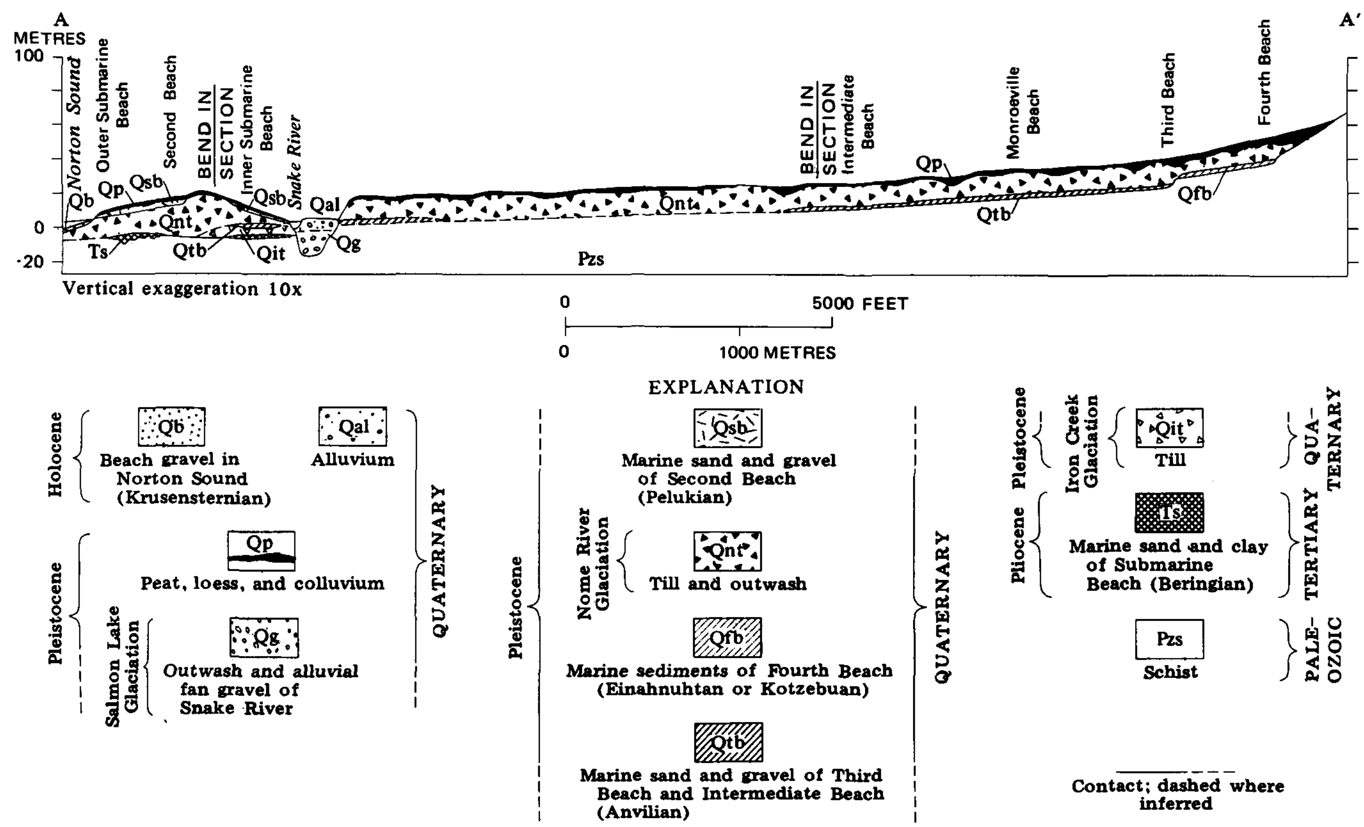


FIGURE 38.—Cross section through coastal plain at Nome (from Hopkins, 1967a). Recent work by Hopkins and Nelson at Nome reveals that drift of the Nome River Glaciation extends 2,500 m offshore, and thick deposits of drift of the Iron Creek Glaciation extend 4,500 m offshore. Also, a Pliocene beach was reported at the depth of 50 m offshore, as well as Pleistocene beaches lying offshore from 1,000 to 7,500 m at depths of 12, 22, and 25 m (Nelson and Hopkins, 1972, fig. 5; see also Tagg and Greene, 1970).

gested that there were two distinct episodes during which sea level was higher than now but not as high as during the later Anvilian transgression. Later work (Hopkins, unpub. data, 1974) indicates the beds of Skull Cliff are no older than middle Pleistocene.

The deposits of Beringian age are considered to be of

Pliocene age (D. M. Hopkins, oral commun., Dec. 10, 1969) (table 1). (See also Tagg and Greene, 1970, figs. 3, 13.) The Beringian transgression beds on St. George Island in the Pribilofs were deposited during the early part of the Matuyama reversed geomagnetic polarity epoch. The pillow lava overlying the youngest part of the Beringian deposits on this island has a potassium-argon date of 2.13±0.06 m.y. (Cox and others, 1966) (table 3).

ANVILIAN TRANSGRESSION

The type locality of the Anvilian transgression is near Nome (fig. 38), and the transgression occurred after the first glacial advance recorded in the area (table 2). Deposits of this transgression can be distinguished from the preceding Beringian and the following Einahnuhtan transgression by their fauna and stratigraphy. The fauna of the Anvilian transgression contains appreciable numbers of extinct species but more closely resembles modern faunas than do Beringian forms.

The deposits at Nome consist of a discontinuous sheet 2–3 m thick of gold-bearing and fossiliferous sandy and gravelly beach and littoral sediments that extend inland 5 km from the present coastline. These sediments are known locally as Third Beach, Monroeville Beach, and Intermediate Beach. In places the deposits lie on drift of the Iron Creek Glaciation of early Pleistocene age, and in most places they are covered by drift of the Nome River Glaciation (table 2).

According to Hopkins (1967a), sediments of the Anvilian transgression perhaps constitute most of the Gubik Formation in that part of the Arctic Coastal Plain that lies inland from the highest beaches of the Einahnuhtan transgression (the middle Pleistocene transgression of McCulloch, 1967). The formation extends 120 km from the sea and is 1–50 m thick. Both Black (1964) and O'Sullivan (1961) described measured sections of the Gubik but have no age control; Hopkins (1967a) considered parts of the measured sections to be of Anvilian age on the basis of fauna and stratigraphy.

Other localities with evidence of the Anvilian transgression include Solomon, 55 km east of Nome, the Pribilof Islands, and tentatively, some of the Aleutian Islands. Sea level probably stood at least 20 m higher than at present and may have been as high as 100 m above present sea level.

The deposits of the Anvilian transgression at Nome were considered for many years to be Pliocene (Dall, 1920; MacNeil and others, 1943); however, new findings show that the Anvilian deposits at Nome were laid down after the Iron Creek Glaciation. A recent discovery of well-preserved Anvilian deposits with a rich fauna (Hopkins and others, 1974) was made on California Creek in western Seward Peninsula. Study of this area suggests that two glaciations preceded the middle Pleistocene Nome River Glaciation, rather than one early Pleistocene glaciation. A shoreline of 90 m on St. George Island lies considerably above fossiliferous Beringian beds and is cut in reversely magnetized volcanic rocks, which belong to the latter part of the Matuyama polarity epoch and which are dated as between 0.7 and 1.9 m.y. old. These lines of evidence suggested to Hopkins (1967a) that the shoreline is of Anvilian age.

EINAHNUHTAN TRANSGRESSION

The Einahnuhtan transgression is represented by a sequence of fossiliferous beach and littoral sediments about 30 m thick at its type locality in the Einahnuhto Bluffs on St. Paul Island in the Pribilofs. These deposits overlie a basaltic flow that has been dated at 280,000 and 380,000 years (Cox and others, 1966) and in turn are overlain by sediments of Kotzebuan age that are capped by a lava flow around 100,000 years old. Sediments of Einahnuhtan age were reported by McCulloch (1967) on the Arctic Coastal Plain. Fourth Beach at Nome may be either Einahnuhtan or Kotzebuan in age.

The name Middletonian transgression was proposed by Karlstrom (1965, 1968) for a high sea-level stand that he believed took place at about the same time as the Einahnuhtan transgression. Hopkins (1967a) did not use this term because he believed the results of radiometric dating of the Middletonian beds are open to question, and the molluscan fauna suggests the beds are Anvilian in age.

The fauna of the Einahnuhtan deposits is essentially modern in aspect. The Einahnuhtan transgression represents an interglacial interval of late middle Pleistocene time.

KOTZEBUAN TRANSGRESSION

The Kotzebuan transgression took place during the interglacial interval that immediately preceded the Illinoian Glaciation (table 3). The type locality is the sea cliffs facing Kotzebue Sound on the west shore of Baldwin Peninsula south of Kotzebue (McCulloch and others, 1965). The deposits are silty clay and peaty silt of deltaic origin. In places they are covered by drift comparable in topographic expression to drift of the Nome River Glaciation at Nome. Sediments of the Kotzebuan transgression are recognized in many other places on the shores of Kotzebue Sound. Sediments of possible Kotzebuan age are recognized on the Pribilof Islands, Cape Prince of Wales, and Nushagak and Kvichak Bays of southwestern Alaska. McCulloch (1967) tentatively assigned some of the marine deposits along the northwest coast to this transgression.

Sea level during the Kotzebuan transgression in Kotzebue Sound was about 21–36 m higher than at present, and the sediments generally lie inland from, and higher than, the well-preserved wave-cut cliffs and ancient beach ridges formed during the following Pelukian transgression (Sangamon).

On western St. Lawrence Island, sediments of Kotzebuan age occur under drift of Illinoian age. A comparison of the mollusk fauna of this age from St. Lawrence Island, Chukotka Peninsula, and Kotzebue Sound suggest that during Kotzebuan time cold Arctic water pushed southward through the Bering Strait, a flow which is reversed from the present northward flow through Bering and Anadyr Straits (Hopkins and others, 1972).

PELUKIAN TRANSGRESSION

The type locality of deposits of the Pelukian transgression is at Nome and consists of a sheet of marine sand and gravel 3–7 m thick, overlying drift of the Nome River Glaciation, and a layer of estuarine sandy silt and black organic clay 2–4 m thick on the south side of the Snake River valley, also lying on drift of the Nome River Glaciation (fig. 38). Pelukian sediments are covered by eolian silt and colluvium of late Wisconsinan and Holocene age and were deposited during an eustatic rise of sea level that coincided with the Sangamon Interglaciation. The Pelukian transgression is represented by well-preserved marine terraces and fairly well preserved marine scarps that lie inland from Holocene deposits.

Deposits of the Pelukian transgression are also present in northern and western Alaska (Sainsbury and others, 1965; Sainsbury, 1967a, b) as well as on St. Lawrence Island and the Pribilofs. The highest terraces along the coast of the Gulf of Alaska between Katalla and Icy Point may have been carved by the sea during the Pelukian transgression. Two distinct Pelukian sea-level stands probably lay between 7 and 10 m above present sea level.

A fossiliferous interglacial beach deposit at South Bight on Amchitka in the Aleutian Islands (pl. 1), 35 m above sea level, yielded bones of Steller's sea cow (*Hydrodamalis*) which were dated as 135,000 years (Gard and Szabo, 1971; Gard and others, 1972). It is thought the deposit in this tectonically active area is Sangamon (Pelukian) in age.

WORONZOFIAN TRANSGRESSION

The marine Bootlegger Cove Clay in the vicinity of Anchorage constitutes type sediments for the Woronzofian transgression as described by Hopkins (1967a, b). The clay contains mollusks and Foraminifera (Hansen, 1965, p. A20) as well as ostracodes (Schmidt, 1963). It is as much as 16 m thick and interfingers laterally with proglacial outwash deltaic sediments. Underlying it is drift of the Knik (early Wisconsinan) Glaciation (table 2); overlying it are outwash sediments and an end moraine of the Naptowne (late Wisconsinan) Glaciation. On the basis of a Th 230 date of 33,000–48,000 years (Sackett, 1958) on mollusk shells from the Bootlegger Cove Clay, Hopkins (1967a) assigned an age of 25,000–35,000 years to the Woronzofian transgression. He believed that sea level then was a few metres lower than at present, as also suggested by Curray (1965).

Because the Bootlegger Cove Clay has evoked widespread interest and controversy both as a stratigraphic unit important in establishing a late Pleistocene chronology for southern Alaska (p. 24, 73) and as the formation responsible for the disastrous landslides that occurred in Anchorage during the 1964 earthquake (p. 73), Ernest Dobrovolny and H. R. Schmoll (oral commun., 1970), in their detailed geologic work in the Anchorage area, attempted to determine the age of the unit more precisely. Four radiocarbon age determinations on shells indicate that the Bootlegger Cove Clay was deposited about 14,000 years ago (Schmoll and others, 1972). A new uranium-series age of 15,200 ± 2,800 years on shells supports the radiocarbon age (Schmoll and others, 1972, p. 1110); thus the previously reported uranium-series dates are apparently incorrect.

As reported by Schmoll, Szabo, Rubin, and Dobrovolny (1972), the Bootlegger Cove Clay particularly the zone with some microfossils, may be interpreted as representing a marine transgression during an early phase of eustatic rise in sea level that followed the maximum extent of glaciation in Wisconsinan time.

Inasmuch as the Bottlegger Cove Clay was designated as the type deposit for the Woronzofian marine transgression, it is clear that this transgression can no longer be considered a mid-Wisconsinan interstadial as defined by Hopkins (1965, 1967a) or a pre-Wisconsinan interglacial event as considered by Karlstrom (1964, 1968). As Schmoll, Szabo, Rubin, and Dobrovolny (1972) suggested, the term Woronzofian should either be redefined as an event of late Wisconsinan age (as already suggested by Morner, 1971) or be discarded to avoid further confusion. In any case, dated deposits elsewhere that have been considered part of the Woronzofian transgression, such as marine deposits at Point Barrow (MacCarthy, 1958; Hopkins, 1967a; Brown, 1965a) or the Karginsky deposits of northern Siberia (Hopkins, 1965; Karlstrom, 1968; Kind, 1967), probably should be reclassified.

Schmoll, Szabo, Rubin, and Dobrovolny (1972) illustrated that other events in various parts of the world, however, can be correlated, on the basis of the new radiometric dates, with the transgression represented by the Bootlegger Cove Clay. Among these are (1) a short-lived warm phase at 14,100 years B.P. recorded in ice cores in Greenland (Dansgaard and others, 1969); (2) deposition of marine "Clay III" at Boston (Kaye, 1961) from which radiocarbon dating of four shell samples yielded an average age of 13,900±300 years B.P. (Kaye and Barghoorn, 1964; Krueger and Weeks, 1966, p. 142; Stuiver, 1969, p. 566); and (3) a sea level stillstand at −38 m in the Bering and Chukchi Seas which probably is between 13,000 and 14,000 years old (D. M. Hopkins, oral commun., April 1972; Creager and McManus, 1967; Sheth, 1971, p. 13).

KRUSENSTERNIAN TRANSGRESSION

Marine deposits of the last 10,000 years or so are from the transgression that resulted from the melting of continental glaciers during the late Wisconsinan and Holocene time. This was named the Krusensternian transgression, from Cape Krusenstern in northwestern Alaska. In most parts of Alaska, deposits of the Krusensternian transgression lie within a few metres of present sea level and are less than 6,000 years old; however, elevated marine terraces in southern and southeastern Alaska that formed 11,000–6,000 years ago are considered to have been carved early in the Krusensternian transgression.

Archeological studies of associated cultural sequences suggest that progradation of the coast began, in different places, from 1,000 to 4,500 years ago. The marine sediments of the vast Yukon-Kuskokwim Delta are in part of Krusensternian age, and many major tributaries were diverted during Krusensternian time. Hopkins (1967a) believed that sea level lay far below its present position until 6,000 years ago and within a few metres of its present position since about 5,000 years ago. This is supported at Barrow by radiocarbon dating of coastal peat (Brown and Sellmann, 1966a) and by geothermal studies by Lachenbruch (1957). However, Moore (1960) suggested from a study of arctic beach ridges that sea level rose about 3 m during the last 5,000 years.

Regressions of the sea during glacial maximums in Pleistocene time exposed many shallow-water offshore areas in the world. One of the most interesting and well-studied areas is the Bering-Chukchi platform, a wide, gently sloping area between the Alaskan and Siberian coasts and now covered by 30–150 m of water (Scholl and others, 1974). Although this area formed a barrier to the migration of marine organisms during glacial times, it also provided a land bridge between Siberia and Alaska for the migration of vertebrates, including man, during these times. This wide, treeless land bridge with a severe Arctic climate existed during major Pleistocene glaciations (Hopkins, 1959b; 1967a, p. 451–484), and the history of the land bridge during the past 35,000 years is summarized as follows: (1) Bridge present more than 35,000 years ago; (2) bridge narrowed and probably severed during middle Wisconsinan time about 35,000–25,000 years ago; (3) land connection restored during late Wisconsinan advances from 25,000 to 12,000 years ago, although land connections may have been drowned occasionally for short periods during the interval; and (4) land bridge drowned by the rising sea between 11,000 and 10,000 years ago.

SUMMARY

The first clear stratigraphic picture of Alaskan marine deposits of Quaternary age now above sea level has appeared over the past decade. Well-preserved beaches and shallow-water sediments from late Pliocene to Holocene age represent marine transgressions during higher eustatic stands of sea level. The transgressions are fairly well dated radiometrically and are given provincial names. Because eustatic sea level changes can be correlated worldwide, they provide an excellent means of tying Alaskan Quaternary events into happenings elsewhere.

VOLCANIC ASH DEPOSITS

Volcanic ash deposits are widespread and abundant in Quaternary sediments throughout southern Alaska from the far west in the Aleutian Islands to southeastern Alaska. Ash deposits are present but less abundant in central Alaska, and are unknown in northern Alaska. An ash layer constitutes an ideal stratigraphic marker and forms the basis of what Wilcox (1965) calls volcanic-ash or volcanic-ejecta chronology. Ash deposits have been useful in solving stratigraphic problems throughout the world; Thorarinsson (1944, 1949) used the name "tephrochronology" for the use of ash deposits in stratigraphic correlations in Iceland.

The ash deposits in Alaska have been noted for years, but despite their potential, little use has been made of them in Quaternary stratigraphy. A technique recently developed for ash identification—accurate, rapid, quantitative chemical analyses of the glass shards of an ash by using the electron microprobe (Smith and Westgate, 1969)—should greatly stimulate use of recognized ash horizons in Quaternary work (Smith and others, 1969).

Potassium-argon methods make it possible to date some ash horizons, especially near the source where the ash may be thick. A radiometrically dated ash horizon is an excellent marker bed when traced from the source into sediments of different origins. The distribution of an ash bed may also be used to infer wind directions at the time of eruption. Such work has already been done in Alaska with the Katmai ash (Curtis, 1969, p. 159) and the White River Ash (Lerbekmo and others, 1968).

SOUTHERN ALASKA

In Southern Alaska, as might be expected of this volcanic area, ash beds are numerous. However, since much of this area was once covered and scoured by glacier ice, almost all the ash layers are postglacial or post-Wisconsinan in age. Many layers have proved useful in providing dates in several areas. Ash horizons have been used by Black to date Anangula, the oldest archeological site in the Aleutians (Black and Laughlin, 1964), estimated to be between 8,000 and 12,000 years old.

Heusser (1960, p. 186–188; table 5) used ash layers as stratigraphic markers in his pollen work along the south and southeast coast of Alaska. An ash layer about 6,700 years old is recorded on Kodiak Island and Kenai Peninsula; an ash layer about 1,200 years old is noted in the Prince William Sound and Icy Cape section.

An ash bed in southeastern Alaska has been dated as about 2,300 years old. This may be the same as the ash layer from Mount Edgecumbe described by Reed and Coats (1942, p. 48–57). McKenzie (1970a) suggested an age of 11,000 years for an ash layer at Glacier Bay thought to be derived from Mount Edgecumbe.

At least four pre-Katmai ash marker beds occur in peat sections at Brooks Lake just northwest of Mount Katmai. The pre-Katmai ash beds are younger than 6,000 years (Heusser, 1963a; Nowak, 1969; L. S. Cressman and D. E. Dumond, unpub. data, 1962).

Miller and Dobrovolny (1959, p. 78) mentioned ash layers 1–10 mm thick in bogs near Anchorage. No mention of age is made except that they are perhaps post-Wisconsinan. Dachnowski-Stokes (1941) also mentioned ash horizons in bogs near Anchorage, as well as elsewhere in Alaska.

The most outstanding ash fall in Alaska during historic time is from the 1912 eruption of Mount Katmai and Novarupta in the Aleutian Range (pl. 1). Griggs (1922, p. 29) estimated that 20×10^9 m^3 of ash was ejected over an area of at least 280,000 km^2. Ash deposits as much as 15 m thick exist in the mountains, and an ash layer 6 mm thick was deposited at a distance of at least 660 km from the mountains.

The tephra described by Griggs (1922) and by Fenner (1920) was restudied in considerable detail by Curtis (1969). This new work shows that nine layers of tephra, each with distinguishing characteristics, can be recognized and that practically all the ejecta came from Novarupta instead of Mount Katmai. This so-called Katmai ash is now a valuable stratigraphic marker and has been recognized by Heusser (1960, table 5) on Kodiak Island and on Kenai Peninsula.

The glaciers adjacent to Mount Katmai were buried under many feet of ash, and the ablation of the ice has been severely hindered since 1912. Except for reconnaissance observations of changes in glacier regimen by Muller and Coulter (1957), the opportunity to study glacier regimen and ablation changes caused by a thick ash deposit of known age has been little used.

One significant application of volcanic ash beds is that of marker beds in firn fields of glaciers in Alaska. It is suggested that the 1912 Katmai ash bed can be identified in the firn of the Greenland icecap (C. C. Langway, Jr., oral commun., 1964).

The application of the new techniques, which permit precise dating and chemical identification of source, may make the isolated small occurences of ash more useful in deciphering Quaternary history in Alaska. There is great potential for the use of recognizable ash horizons in archeology, paleopedology, and palynology. This is especially true in southern Alaska, where volcanic ash is still being deposited. On July 9, 1953, a 6-mm layer of volcanic ash from Mount Spurr was deposited on the city of Anchorage (Juhle and Coulter, 1955; Wilcox, 1959, fig. 63). The ash fall partly blocked the sunlight.

SEWARD PENINSULA

An area of about 1,000 km in northern Seward Peninsula near Cape Espenberg is overlain by volcanic ash. The ash is derived from several large maars the youngest of which, Devil Mountain, erupted in late Wisconsinan time, near the end of the last episode of loess deposition in the area (D. M. Hopkins, unpub. data, 1968).

CENTRAL ALASKA

The oldest volcanic ash layers in central Alaska are known from gold mining exposures in the central part of the Yukon-Tanana Upland, especially over a large area in the vicinity of Fairbanks (Péwé, 1952a). Except for the White River Ash of Lerbekmo, Hanson, and Campbell (1968) in east-central Alaska, no source area has yet been identified. Ash horizons, from Illinoian to Holocene, have been used in stratigraphic work near Fairbanks. The oldest and thickest deposit is the Ester Ash Bed (fig. 39; table 3) (Péwé, 1955, p. 713; 1965a, fig. 1–8), a 15-cm-thick white vitric ash bed of at least early Illinoian age; it lies near the base of the oldest known loess deposit. A white vitric ash bed (Péwé, 1952a, p. 86), 1–2 cm thick, thought to be late Illinoian in age, is exposed in many placer cuts (fig. 29). In all exposures the ash has been contorted by solifluction movement and block slumping (Péwé, 1975). A 1-cm-thick white ash layer is known from Wilber Creek (Péwé, 1952a, p. 156), where it is thought to be less than 4,200 years old. A 1-cm-thick white ash layer, dated as 14,000 years old by bracketing radiocarbon dates, is exposed along the Chatanika River, 40 km north of Fairbanks, and has

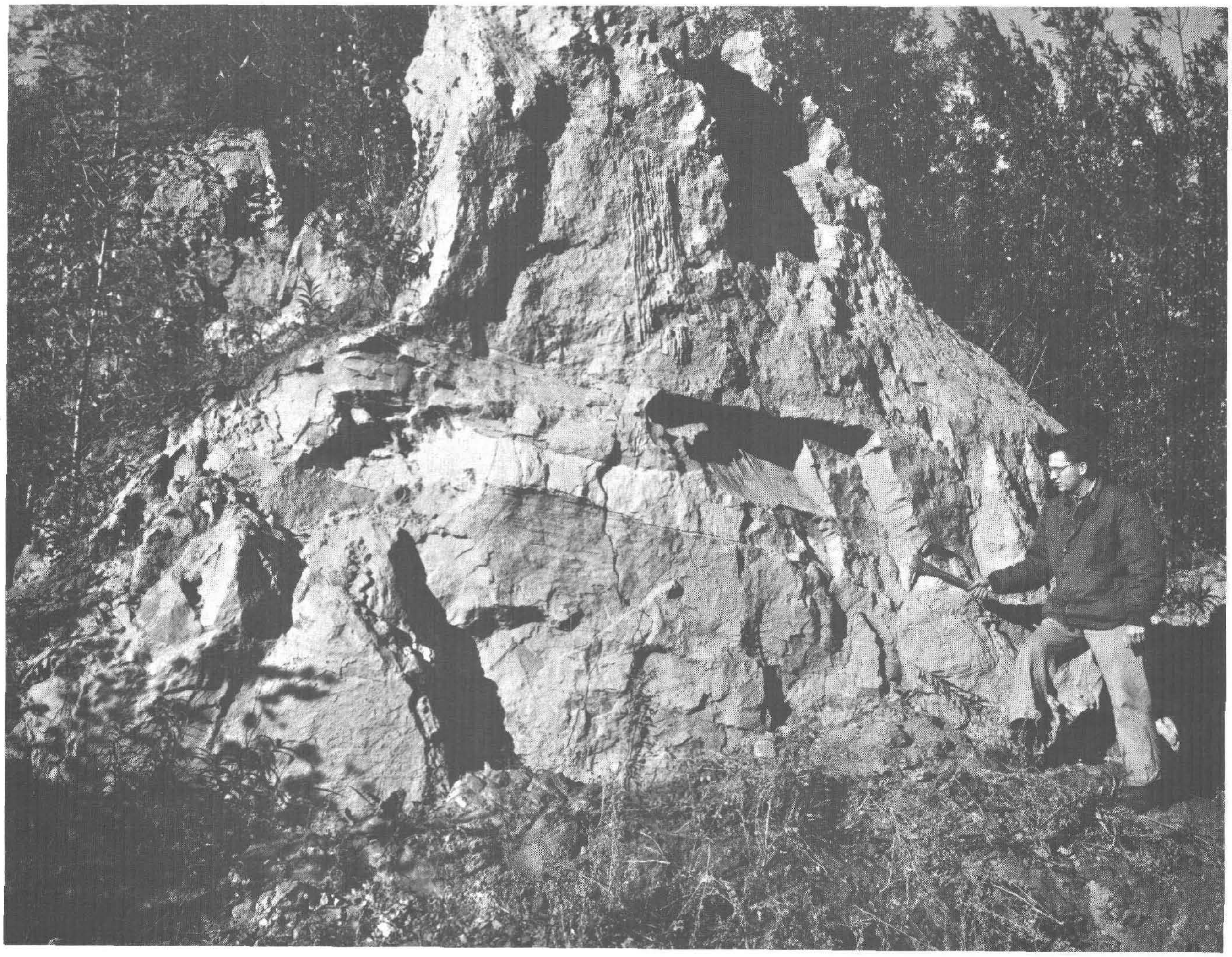


FIGURE 39.—Ester Ash Bed of Illinoian age in Fairbanks Loess, exposed near base of mining cut between Ester and Cripple Creeks 16 km east of Fairbanks, Alaska. Photograph No. 354 by T. L. Péwé, August 23, 1948.

been named the Chatanika Ash bed (Péwé, 1975). The ash layer is characteristically distorted by ice wedge growth (Péwé, 1965a, fig. 1–15). The stratigraphic position of the ash layers seems fairly well authenticated (Péwé, 1975); electron microprobe analysis is now underway to refine these interpretations and correlate these ash layers with other exposures.

A distinctive 6-mm-thick white ash bed, widespread in the lower Delta River area, is well exposed along the Delta River and Jarvis Creek and has been termed the Jarvis Ash Bed (Péwé, 1965c, p. 63–66). The ash layer is exposed in many loess sections and is dated as 2,000–4,000 years old (Péwé and Holmes, 1964; Reger and others, 1964).

The most remarkable ash fall recorded in Alaska is a light-colored pumiceous ash described by Capps (1916a) in the upper White and Yukon River valleys and named the White River Ash by Lerbekmo, Hanson, and Campbell (1968). It was termed an andesitic pumiceous ash by Knopf (Moffit and Knopf, 1910, p. 43–44) and a rhyodacitic or dacitic ash by Berger (1960). The possible source of the ash was Mount Natazhat, at the east end of the Wrangell Mountains approximately 7 km west of the Alaska-Canada border (Lerbekmo and others, 1968, p. 284). Capps (1916a) estimated that about 40×10^7 m^3 of ash was deposited over an area of more than 360,000 kr ^{2}. Ash beds as much as 90 m thick are noted near the source. The ash thins to less than 0.5 m in 30 km and then spreads a thin blanket 1–25 mm thick for 500–700 km north and east in a bilobate pattern (Lerbekmo and others, 1968; Lerbekmo and Campbell, 1969). The thin seam of white ash is widespread in east-central Alaska and western Yukon Territory and has been reported since 1885 (Schwatka, 1885; Dawson, 1889; Hayes, 1892; Johnson, 1946; Bostock, 1952; Hansen, 1953; Hanson, 1965). The ash lies generally just beneath the vegetation mat, except where peat or loess accumulation has been rapid.

Capps (1916a) estimated the age of the ash fall to be 1,400 years by counting growth rings of spruce trees that produce successively higher lateral roots as peat accumulated over the ash with consequent rise of the permafrost table. The accuracy of this ingenious dating method has been borne out inasmuch as Fernald's (1962) radiocarbon data indicate that the ash layer is about 2,000 years old, and Stuiver, Borns, and Denton (1964) radiometrically dated the ash fall at 1,425 years in adjacent Canada. Recent work (Lerbekmo and others, 1973) suggest the ash fall may be more complex than previously realized. There may be three layers closely spaced in time, with the youngest about 1,000 years old.

The White River Ash has proved useful in archeological studies, studies of Holocene glacial advances (Borns and Goldthwaite, 1966, p. 615), and investigations on the deposition rate of loess and peat.

Heusser (1960, table 5, p. 188) believed that a thin ash horizon in peat bogs from Kodiak Island to Icy Cape near Mount Saint Elias is correlative with the ash bed described by Capps (table 3). He placed the age of the ash, in absence of radiocarbon dates, at about 2,500 years. The work of Lerbekmo, Hanson, and Campbell (1968), however, showed that the White River Ash did not drift along the southern coast. The ash described by Heusser, therefore, has a different source, and its age estimate is less tenable.

In addition to these stratigraphically controlled ash horizons in central Alaska, there are many other ash beds, some of them mentioned in the literature. Fernald (1960, p. 251) and Capps (1935, p. 87) reported a thin ash layer near the surface in the upper Kuskokwim region. Florence Weber (oral commun., 1961) noted a lenticular pocket of white volcanic ash 30 cm thick in Quaternary sediments along the lower Koyukuk River.

FLORA

The present-day vegetation of Alaska consists of three major types: the coastal Sitka spruce–hemlock forests of southeastern and south-central Alaska, the interior white spruce–birch forest of central Alaska, and the treeless tundra of western and northern Alaska (Sigafoos, 1958) (fig. 40).

The coastal forests extend a few tens of kilometres inland along the coast of southeastern and southern Alaska westward to Cook Inlet and northeastern Kodiak Island. In addition to the Sitka spruce *(Picea sitchensis)* and western hemlock *(Tsuga heterophylla)*, the coastal forest includes mountain hemlock *(Tsuga mertensiana)*, Alaska cedar *(Chamaecyparis nootkatensis)*, lodgepole pine *(Pinus contorta)*, and Douglas fir *(Pseudotsuga menziesii)*, and smaller areas of Alpine fir *(Abies lasiocarpa*, Pacific silver fir *(Abies amabilis)*, and western redcedar *(Thuya plicata)* (Sigafoos, 1958; Heusser, 1960).

The interior forest is distributed today through most of central Alaska north of the coastal mountains and south of the Brooks Range. In addition to the white spruce (*Picea glauca*) and birch (*Betula papyrifera*), the common species are black spruce (*Picea mariana*), balsam poplar (*Populus balsamifera*), aspen (*Populus tremuloides*), larch (*Larix larician*), and willow (*Salix sp.*) (Sigafoos, 1958; Hiltén, 1941–48). Various species of alder (*Alnus*) grow in both forest and tundra regions.

The tundra is found beyond the limits of forest in central and western Alaska and elsewhere in highland areas above tree line. Tundra covers approximately 25 percent of Alaska and is a mosaic of many different types of vegetation, some of which are limited to either southern or northern tundra regions (Griggs, 1936; Spetzman, 1959; Britton, 1958b; Wiggins, 1962).

The distribution of vegetation in Alaska falls into two large phytogeographic areas: the glaciated and the unglaciated regions (fig. 6). The glaciated areas have a floral history of repopulation of the land by either tundra or forest during the last 1,000 to tens of thousands of years. The unglaciated part of Alaska constitutes 50 percent of the State and represented refugia for plant and animal life during glacial maximums (Hultén, 1937; Heusser, 1960). In the unglaciated areas, especially of central and western Alaska, the vegetation history records the shifting of tree line and the change from one type of tundra vegetation to another.

In general, the changes in vegetation from glacial to interglacial times in Alaska are small compared with the great changes in temperate latitudes (Sears, 1935, 1938). Yet, recognizable changes in the vegetation did occur in Alaska throughout Quaternary time.

The record of Quaternary vegetation is based largely on palynology. In the Arctic, however, palynology is a blunt instrument because the tundra pollen diagram is much simpler and less informative than forest records of temperate regions (Colinvaux, 1967a). Information concerning vegetation of Wisconsinan age or earlier is limited mainly to deposits of the Yukon-Tanana Upland and to terrestrial and marine sediments of western and northern Alaska. Studies of post-Wisconsinan vegetation are available for some parts of Alaska, particularly of the southern and southeastern coastal areas.

EARLY TO MIDDLE PLEISTOCENE

The floral record in Alaska earlier than about Yarmouth(?) time is scanty and limited to western Alaska, particularly Seward Peninsula. Because the classical midwestern glacial terminology of North America has not been dated, the early Quaternary deposits in Alaska cannot be directly related to it. However, the deposits

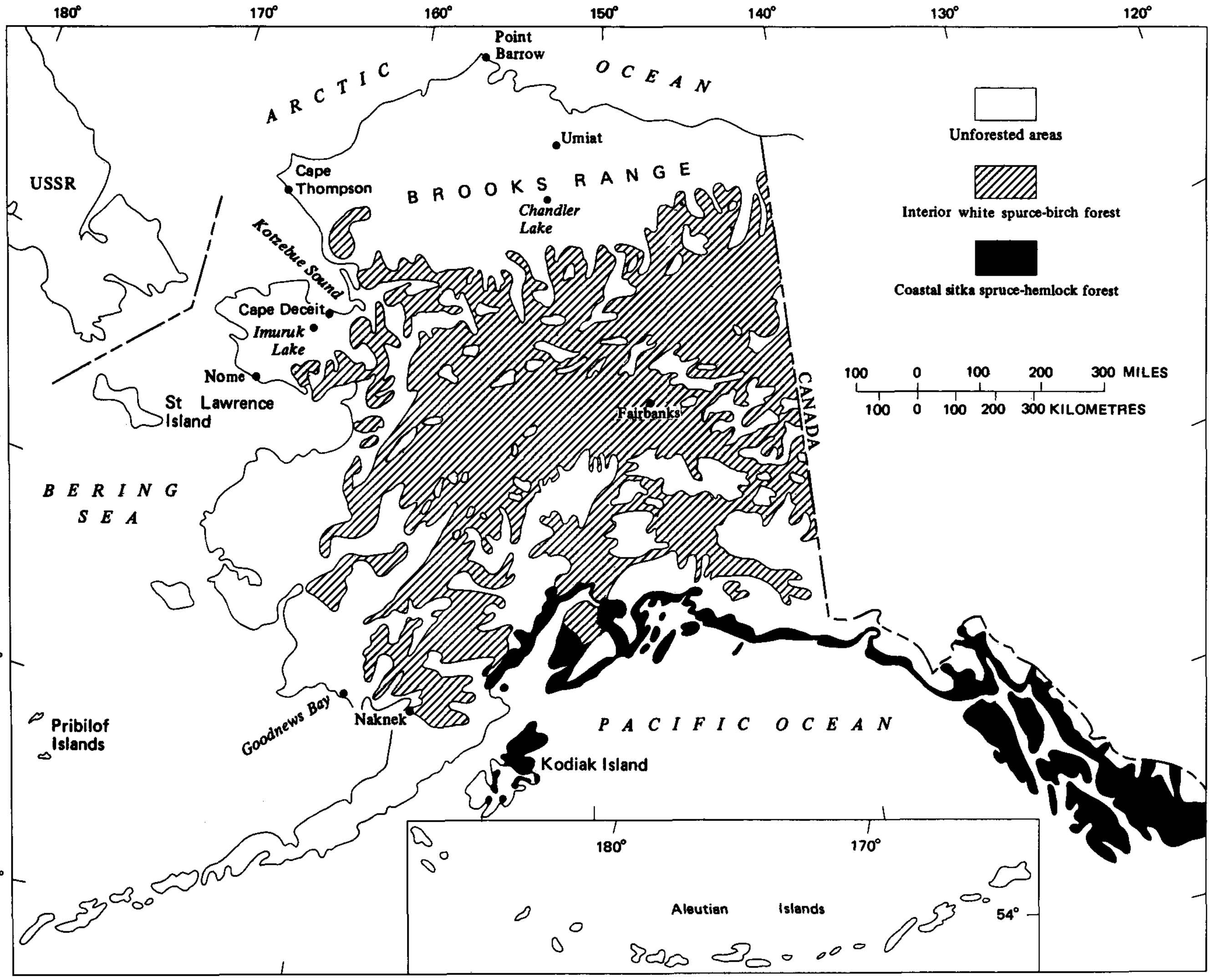


FIGURE 40.—Generalized map of major forest types and unforested areas in Alaska. After Sigafoos (1958) and Hopkins (1959a).

can be related to named Alaskan marine transgressions or glaciations.

The earliest record of flora in Alaska important to the Quaternary history is the pollen and wood preserved in deposits of the Beringian transgression (late Pliocene) in western and northern Alaska and the Aleutian Islands (Hopkins, 1967a). *Picea* and *Pinus* were reported on the Arctic coast, near Nome, and on Tanaga Island in the Aleutians; *Larix* also occurs in the coastal and Nome areas (Hopkins, 1965). *Abies, Tsuga,* and *Carya* that were reported in Beringian deposits at Nome (Hopkins, 1965) probably represent pollen reworked from older beds (Wolfe and Leopold, 1967, p. 197–198; Hopkins, 1967a, p. 61) a few miles off the present coast (Nelson and Hopkins, 1972). It is noteworthy that in Beringian time *Picea* and *Larix* were several hundred kilometres beyond their present limits, and *Pinus* was many hundreds of kilometres beyond its present limit.

The lowest part of a lacustrine deposit in the Yukon Flats contains a pollen flora of pine, spruce, alder, birch, hemlock, fir, and hickory (Williams, 1962, p. 304) and may be late Pliocene or early Pleistocene in age.

The record of pre-Yarmouth(?) Quaternary flora is limited to the detailed study made by Guthrie and Matthews (1971) on a complex sequence of vegetation-rich silts at Cape Deceit near Deering on the south shore of Kotzebue Sound. The sediments are rich in fossil vertebrates and flora, peat, pollen, and tree macrofossils. The vertebrate study was reported (Guthrie and Matthews, 1971) along with comments about the vegetation; work on description and interpretation of the flora is not yet finished. This study reported the most complete sequence of Quaternary sediments in Alaska, from perhaps earliest Pleistocene to and including the Holocene. The section is divided into three formations: the Cape Deceit Formation, early Pleistocene to and including Iron Creek Glaciation [Gunz I(?)-Gunz I/Gunz II(?)-Gunz II(?)]; Imnachuk Formation [Mindle/Gunz II(?)-Mindel(?)-Einahnuhtan or Anvilian transgression]; and Deering Formation [from Mindel/Riss(?) (Kotzebuan transgression) to present].

Guthrie and Matthews (1971) believed that on the north side of eastern Seward Peninsula during a very early interglacial(?) period thought to be Waal(?) (Gunz I/Gunz II) a forest tundra existed and conifer tree line advanced as far as Cape Deceit. *Larix* was sparsely distributed in the region. This environment was followed by a period with a more severe tundra environment than now and with the formation of nonwoody peat. This is thought to correlate with the time of the Iron Creek Glaciation (tables 2, 3).

No vegetation record is present for the next interglaciation, which may be either the Einahnuhtan or Anvilian, but a peat deposit representing a tundra environment is thought to be equivalent to a glaciation (Mindel?).

YARMOUTH(?)

Three types of evidence, each from different areas, are available to indicate an interglaciation prior to the Illinoian Glaciation, the time of the Kotzebuan transgression (Yarmouth?): marine and nonmarine beds from Kotzebue Sound, pollen from lake deposits in central Seward Peninsula, and wood at the base of loess near Fairbanks.

Hopkins (1967a) stated that plant remains are abundant in beds of the Kotzebuan transgression in the Kotzebue Sound area but as yet have received little study. Wood of *Pinus* and *Populus* is known, and *Pinus* pollen in silt is recorded. The presence of pine should be investigated further because studies at nearby Cape Deceit and Imuruk Lake reveal spruce but no pine at that time. Pine does not extend to central and western Alaska today. Guthrie and Matthews (1971) indicated that at Cape Deceit during the time of the Kotzebuan transgression(?) (Yarmouth?) the area was forested and spruce tree line advanced to at least that position (fig. 40).

Colinvaux (1962, 1963) studied the pollen of an 8-m-long core from sediments of the bottom of Imuruk Lake in central Seward Peninsula (fig. 40). He found a record of changing tundra types consisting of three periods when vegetation was similar to now, alternating with periods when vegetation was different (colder). Because of an anomaly in the radiocarbon dating, the record can be interpreted as representing either the last 22,000 years or the span of time from a pre-Illinoian interglaciation (Yarmouth?) until the present. Colinvaux favored the latter interpretation, and he (1967a) did not list the "anomalous" radiocarbon dates, although he stated that there was no radiocarbon support for the lower part of the core. Black (1966) stated that the use of the lower part of the core might be meaningless.

This controversy over the interpretation of the lower part of the core may now be solved by Colinvaux (oral commun., Oct. 24, 1969) and Colbaugh (1968). A core from a lake on St. Paul Island in the Pribilofs revealed the same type of anomalously young dates below older dates (Colinvaux, 1967c) as does the Imuruk Lake core. The young dates are not from organic mud but from sandy layers that permitted ground water to flow down the side of the basin under the mud plug and to enrich the sand with modern carbon and so produce the anomalously young radiocarbon ages. This idea of ground-water flow was confirmed with tritium tracers.

The anomalously young dates in the first Imuruk Lake core are also from sandy sediment permitting ground-water flow. Since publication of the original data (Colinvaux, 1964a), other cores collected from Imuruk Lake by Colinvaux have been analyzed. Core II and core IV were analyzed by Colbaugh (1968, figs. 5, 6), including radiocarbon dating, and her work completely supports Colinvaux's interpretation.

Accepting Colinvaux's later interpretation, the pollen record from Imuruk Lake indicates that during a pre-Illinoian interglacial time (Yarmouth?), the vegetation of the area was similar to now, with much tundra flora of cotton grass (*Eriophorum*), dwarf birch (*Betula nana*), and alder (*Alnus*) present. The spruce tree line was not far distant.

In central Alaska at the junction of Engineer Creek and Dawson Cut, as well as at Eva Creek (fig. 29), Sheep Creek, and Ready Bullion Bench, there is exposed in mining cuts at the base of the Illinoian loess a bed of large white spruce stumps (22 cm in diameter) in growth position (fig. 41) and prostrate spruce logs 3 m long and 30 cm in diameter (Péwé, 1952a). These plant remains appear older and more weathered than wood in the overlying Illinoian and Wisconsinan frozen loess. This wood is smashed, flattened, and iron stained. These stumps date from a time of a well-developed forest, a time that may be part of the interglaciation just prior to the deposition of the Illinoian loess. This forest bed has been termed the Dawson Cut Formation (Péwé, 1975). The wood at the base of the loess and on the underlying tightly cemented iron-stained gravel at Sheep Creek was identified as spruce and that from Ready Bullion Bench as alder (R. C. Koeppen, written commun., Sept. 28, 1967).

ILLINOIAN

The record of flora in Alaska is still scant in deposits of Illinoian age. Information is restricted to data from Seward Peninsula and Fairbanks. The record of the flora for the Illinoian interval of the Imuruk Lake core (Colinvaux, 1967a) shows that there was an Arctic grass tundra with little or no alder. Dwarf birch was

FIGURE 41.—White spruce stumps rooted in soil and gravel at base of loess of Illinoian age, Eva Creek, 16 km west of Fairbanks, Alaska. Large stump was submitted to three different laboratories for radiocarbon dating and was beyond the range of dating—University of Michigan, 18,000 years; University of Chicago, >20,000 years; Lamont Geological Observatory (1–137X), >28,000 years. Photograph No. 471 by T. L. Péwé, September 18, 1949. (See fig. 29.)

present but less common than during interglacial times. Sage (*Artemisia*) was more common than now or in an interglacial period. The tree line was not close. At Cape Deceit (Guthrie and Matthews, 1971) during Illinoian time, a tundra environment was present.

Pollen analysis of peat collected by the writer in Illinoian loess in valley bottoms of the Fairbanks area indicates that the tree pollen content was conspicuously and significantly lower than today, indicating a relatively sparse tree cover. No pollen of the Gramineae was present (E. S. Barghoorn, written commun., May 1, 1957). A lowering of tree line to the level of the creek valley bottoms requires a drop of 450–600 m, which suggests that the climate was colder than now. A much more restricted forest area is also borne out by the greater number of grazers (mammoth, bison, horse) than present (Péwé, 1952a; Péwé and Hopkins, 1967).

In the Fairbanks area today, the boreal forest is well developed, and tree line is approximately 600–700 m above sea level. Above this elevation, the rounded hills are clothed in a tundra rich in dwarf birch and alder as well as willow. Two contemporary peat samples collected from creek valley bottoms by the writer contain tree pollen percentages of 53 and 33 percent. The tree pollen percentage is entirely spruce except for three or four erratic pine pollen grains. The nonarboreal pollen in the samples is 35 percent shrubs and 10 percent herbs. Pollen of the Gramineae is less than 1 percent (E. S. Barghoorn, written commun., May 1, 1957).

SANGAMON

The only records of known Sangamon vegetation reported from Alaska come from three localities on Seward Peninsula, one from the Baldwin Peninsula, two from St. Lawrence Island, and several in the Fairbanks area. Hopkins, MacNeil, and Leopold (1960, p.53) showed from pollen[3] and from spruce logs found in estuarine sediments near Nome that tundra vegetation prevailed and that the forest was near at the time of the Sangamon maximum. In late Sangamon time, the tree line reached the Snake River valley (pl. 1). Spruce trees were 50–80 km west of their present limit in the Kotzebue Sound and Nome areas (Hopkins, 1965). Locally, there was tundra with dwarf birch, and alder bushes were nearby. The pollen spectra suggest (Colinvaux, 1967a, p. 217) that during at least part of Sangamon time, the vegetation near and at Nome was comparable to that of the present.

Colinvaux (1962) studied pollen from the Imuruk Lake core in central Seward Peninsula and reported that the vegetation was a sage-alder tundra similar to the present conditions and that tree line was near.

At Cape Deceit (Guthrie and Matthews, 1971) a tundra environment existed with the formation of nonwoody peat. From two adjacent localities on St. Lawrence Island in the Bering Sea (pl. 1), Colinvaux (1967b) cautiously suggested that pollen analyses show that the local tundra, at a time thought to be Sangamon, was comparable to the modern tundra. All evidence so far indicates that the Alaskan side of the Bering Sea had a vegetation comparable to that of the present.

McCulloch, Taylor, and Rubin (1965) reported that in the Kotzebue area in Sangamon time, the forest had advanced westward into areas that are now treeless tundra. Stumps more than 42,000 years old rest on Illinoian till in this area of the Baldwin Peninsula.

In all the mining-cut exposures in the Fairbanks area, the contact between the retransported silt of Wisconsinan age and the underlying loess of Illinoian age (fig. 19) is represented by a poorly formed to well-formed topographic bench. In sections in Eva Creek (fig. 29), Dawson Cut, Gold Hill and elsewhere, the bench is conspicuous because of a forest layer of rooted stumps and prostrate logs of birch and spruce as much as 25 cm in diameter (Péwé, 1952a). It is very likely that in most localities this concentration of tree remains represents forest of Sangamon time. These concentrations occur at the correct stratigraphic horizon and may represent the return of the forest to the area. Samples collected by the writer from the lower part of the Wisconsinan deposits in the area have all been older than 35,000 years. A sample from the possible Sangamon forest bed on Eva Creek was dated as more than 56,900 years (Matthews, 1968b, p. 207, 1970). The forest bed is now termed the Eva Formation (Péwé, 1975). The forest environment was probably similar to that now.

[3]Samples are from a basal peat on the Second Beach (Pelukian transgression) (table 7) at Nome. Analyzed by E. B. Leopold.

WISCONSINAN

Information needed to reconstruct the vegetation pattern in Alaska during Wisconsinan time comes from two broad areas: the tundra zone adjacent to the Bering Sea, including Point Barrow, and the forested interior near Fairbanks.

From the far west, most of the information is from pollen analysis, and only one record covers the entire Wisconsinan time, the other being mainly the end of Wisconsinan time. In central Seward Peninsula, Colinvaux's (1962) pollen analysis indicated that from early Wisconsinan time until the end of Wisconsinan time, an Arctic herbaceous tundra was present, Arctic sage was relatively common, but dwarf birch was absent or scarce. Tree line was not near. At Nome, analysis of pollen by Leopold (Hopkins and others, 1960) indicated that at least near the end of Wisconsinan time (13,600 years ago), a grass-sage tundra existed. Tundra also existed at Cape Deceit (Guthrie and Matthews, 1971).

Cores from Kotzebue Sound and Ogotoruk Creek to the north contain pollen representing a time near the end of the Wisconsinan. Heusser (1963b) concluded that the bottom of the core from Ogotoruk Creek represents the late glacial period and a tundra with little birch. The core from Kotzebue Sound indicates that 12,000 years ago a tundra with dwarf birch was already present (Colinvaux, 1967a, p. 221).

To the far north at Point Barrow, a peat sample dated at 25,000 years old (Brown, 1965a) lacks alder and birch, and Colinvaux (1967a) believed an herbaceous tundra prevailed in the Arctic Coastal Plain at least in that part of the Wisconsinan glacial episode. Pollen from St. Lawrence Island suggested that in Wisconsinan time a cold Arctic tundra existed in which grasses and sages predominated (Colinvaux, 1967b). A peat sample from near Goodnews Bay collected by Hopkins from a drained kettle lake in an end moraine of late Wisconsinan age was analyzed for pollen by E. B. Leopold. The oldest sample is 11,500 years old and contains pollen of an herbaceous tundra. Colinvaux (1967a) stated that even at the south margin of the Bering land bridge, the Wisconsinan tundra vegetation was comparable to that in the Bering Strait area in late Wisconsinan time.

In central Alaska the greatest source of information concerning the flora of Wisconsinan time is from the exposures made by large-scale placer gold mining in the perennially frozen, retransported, organic-rich silt (fig. 20), termed the Goldstream Formation (Péwé, 1975). The silt contains abundant layers, lenses, and isolated pods of plant remains. Peat beds as much as 2 m thick are present, and frozen nests and seed caches of rodents occur (Wilkerson, 1932, p. 15; Chaney and Mason, 1936, p. 14; Mertie, 1937, p. 193; Péwé, 1952a, p. 177; 1966c, fig. 3). Some forest beds are present as well as beaver dams; wood from one dam was dated at 13,600 years B.P. (Péwé, 1975). Wood remains include alder (*Alnus*), spruce (*Picea*), willow (*Salix*), cottonwood (*Populus*), and birch (*Betula*). The logs and limbs throughout the formation are fresh, and some have pieces of bark. Wood is not nearly so abundant, however, as in the overlying post-Wisconsinan silt deposits. In general, more megafossils occur in central Alaska than in the coastal tundra areas because of the existence of small patches of forests in valley bottoms in the interior.

Peat samples collected by the author from different stratigraphic horizons in the Wisconsinan retransported silt (Goldstream Formation) (fig. 20) in the Fairbanks area were studied by E. S. Barghoorn. He reported that the samples from Fairbanks Creek are statistically and significantly low in tree pollen, indicating a very sparsely forested or almost treeless area. The samples from this creek valley indicate that the tundra was rich in alder and dwarf birch with little grass. Pollen from a thick peat bed near the base of the Wisconsinan retransported silt in Ester Creek indicates that coniferous trees were absent (E. S. Barghoorn, written commun., Nov. 23, 1954). The vegetation of the valley bottoms at the time was a grass-sedge-moss-*Equisetum* cover associated with small stands of dwarf birch and willow. Such evidence indicates a lowering of tree line by 600 m in Wisconsinan time. The most thorough study of pollen from the frozen silts of Eva Creek is by Matthews (1970), who reported conclusive evidence that spruce woodlands were almost totally eliminated from interior Alaska in Wisconsinan time.

Among the vertebrate fossils of the retransported silt are *Citellus undulatus, Lemmus siberius, Dicrostonyx torquatus,* and mountain sheep (*Ovis nivicola*). These mammals do not live in the Fairbanks area today but live near or above tree line or on the tundra. Such mammal distribution also suggests that tree line was 500–600 m lower in the Fairbanks area during Wisconsinan time. The restricted forest area is also shown by the high percentage of grazers at this time (Péwé, 1952a; Guthrie, 1968a).

Repenning, Hopkins, and Rubin (1964) clearly demonstrated lowering of tree line by 500 m in the Tofty area, 130 km to the west of Fairbanks, by the presence of tree-line fauna living in valley bottoms in late Wisconsinan time.

All specimens studied by E. S. Barghoorn or by Rolland Brown of the U.S. Geological Survey or megafossils observed by the writer and those reported by Chaney and Mason (1936) and Chaney (1938, p. 390–391) are of the species living in the Fairbanks area today. Although the flora of the Wisconsinan time in central Alaska contains the same species as today, forest cover was greatly restricted, perhaps limited to major valley bottoms.

Hopkins (1970; 1972) put forth the idea that a treeless vegetation prevailed in central Alaska during "full glacial" time and that the spruce persisted in a refugium somewhere in southwestern Alaska until 6,000–8,000 years ago. The writer believes that such a source is not necessary, at least for central Alaska, because scattered spruce probably persisted in isolated creek bottoms in central Alaska, inasmuch as spruce wood occurs in the frozen retransported silt of Wisconsinan age in the Fairbanks area (fig. 20). Spruce wood in beaver dams, isolated sticks, and rooted stumps have been dated as 13,600, 10,500, 9,685, 8,930, 8,500, and 7,500 years old (Péwé, 1975). Dates of 20,000–30,000 years are also known on small sticks, but the species of wood is not known.

POST-WISCONSINAN

The post-Wisconsinan vegetation record is, as might be expected, the most clearly and thoroughly documented part of the floral history. Evidence of the postglacial thermal maximum or hypsithermal interval is recognized from pollen spectra in northern and northwestern Alaska (north of the Brooks Range). However, from central and western Alaska the record is mainly that of forest return or expansion about 10,000 years ago.

The first pollen record of postglacial time in northern and western Alaska was that by Livingstone (1955) from a 5-m-long core taken from Chandler Lake (pl. 1) in the Brooks Range (fig. 40). He set up a threefold chronology that has been used as a general guide throughout northern and northwestern Alaska. The chronology, which is supported by pollen analyses of the radiocarbon-dated core of valley fill from the unglaciated part of the North Slope near Umiat (Livingstone, 1957) (table 3), shows that an herbaceous tundra was present 8,300 years ago but at 7,500 years ago yielded to a tundra in which dwarf birch was very abundant. About 5,800 years ago, alder became very common, and Livingstone stated the alder maximum is climatically equivalent to and contemporaneous with the thermal maximum. Today, the tundra in the Umiat region is characterized by alder and willow (Churchill, 1955).

Good support for the advance of alder north of the Brooks Range is from megafossils of alder found by Tedrow and Walton (1964) in the upper Killik valley (pl.1). Alder leaves 5,650 years old preserved in the frozen ground 13 m beneath the surface led Tedrow and Walton to believe the climate of the area was warmer 5,000 years ago than now, inasmuch as alder does not grow there today.

Work by Colinvaux (1964b, 1965, 1967a, p. 212–213) on samples collected by Brown (1965a, 1969b) at Point

Barrow revealed that from 8,000 to 9,500 years ago a grass-sage tundra was predominant. A stratigraphic hiatus covers 4,000 years, and the last 5,000 years records a higher alder content and vegetation similar to that now. Colinvaux (1967a, p. 213) stated that no broad change in the vegetation record is detectable in the pollen record of the last 5,000 years. However, because of the 4,000-year gap, it cannot be demonstrated whether an alder maximum, indicating the thermal maximum like that recorded at Umiat, was present in the Barrow area.

A pollen analysis by Heusser (1963b) from peat deposits at Ogotoruk Creek about 500 km west of Chandler Lake records a change from an herbaceous tundra in early postglacial time to a tundra in which birch was common. Heusser showed that the record is similar to the early part of the Livingstone chronology determined in the Brooks Range farther east. Schweger (1971) carefully examined a detailed stratigraphic section on the Kobuk River near Onion Portage representing the last 12,000 years and reported no record of a thermal maximum, but rather a gradual change from tundra vegetation to spruce cover.

For an area near Nome, E. B. Leopold (in Hopkins and others, 1960) analyzed pollen that indicates that a threefold spectrum similar to that of Livingstone is present but that the sequence started earlier. Herbaceous tundra (zone I) was dated as 13,000 years B.P. and zone II (birch tundra) as 10,000 years B.P. Zone III is the alder tundra that began 3,500 years ago. Colinvaux (1967a) recognized these three zones from Imuruk Lake but indicated that they span a much longer period of time, with only the sage alder tundra being postglacial.

In the late forties, Hopkins determined from logs and stumps exposed in mining excavations that about 8,000–10,000 years ago the spruce forest advanced westward from its present limits on the Seward Peninsula (Hopkins an Giddings, 1953). This evidence, plus additional findings of alder, birch, and poplar logs (8,500–9,500 yr) on Seward Peninsula and Baldwin Peninsula (McCulloch, 1967), indicates that the forest spread westward in response to the warming that ended the Wisconsinan Glaciation. Colinvaux (1967a) commented that this temporary westward extension of tree line has not yet been noticed in pollen diagrams, perhaps because pollen work in the tundra or near tundra areas is not a delicate enough measurement.

To the south, on St. Paul Island of the Pribilof Islands, Colinvaux (1967c) noted that the pollen record shows that vegetation similar to the present has existed for the last 9,000 years. From Goodnews Bay, pollen analyses reveal that in the last 10,500 years a threefold spectrum of herbaceous, birch, and alder tundras parallel to the three-zone sequence at Chandler Lake existed.

The record of post-Wisconsinan vegetation distribution on the east side of the Bering land bridge from Point Barrow to Goodnews Bay was summarized by Colinvaux (1967a). Heusser (1960) showed by detailed pollen analyses and radiocarbon dating that from 10,000 to about 8,000 years ago tundra was present on the south coast and that pine forests were present in the southeast (table 3). From about 8,000 to 3,500 years ago, Sitka spruce and hemlock predominated in both areas. In the last 3,500 years hemlock predominated in both areas. In the last 3,500 years hemlock and Sitka spruce have been luxuriant on the southeast coast. The shifting of forest areas with variation of glacier extents in the southeastern part of Alaska in the last 8,000 years was chronicled by Cooper (1923, 1931, 1937, 1942).

A few isolated pollen studies have been made elsewhere in Alaska. For an area near Naknek on the Alaskan Peninsula, Heusser (1963a) showed that in postglacial time alder pollen predominated over birch pollen in the tundra until 7,000 years ago. Birch pollen reached a maximum between 7,000 and 3,000 years ago and then declined in percentage. Spruce migrated from the interior of Alaska (fig. 40) within recent centuries. Little is known of the tundra changes in the Aleutians during Quaternary time, but Anderson and Bank (1952) indicated pollen analyses will be fruitful.

In interior Alaska during post-Wisconsinan time, the birch–white spruce forest expanded rather rapidly around 10,000 years ago from valley bottoms to reclothe the hills as tree line rose 450 to 600 m higher than it was during Wisconsinan time. In the Fairbanks area, retransported organic-rich silt (fig. 20), deposition of which began 10,000 years ago (Péwé, 1975) and is still going on, has abundant plant remains, including rooted stumps. This silt is termed the Ready Bullion Formation (Péwé, 1975). Peat beds 25–30 cm thick can be traced for 25 m. Taber (1943, p. 1538) was the first to recognize the unconformity between the Wisconsinan and Holocene sediments when visiting the Fairbanks area in 1935; however, he attributed it to erosion during Yarmouth interglacial time.

Matthews (unpub. data, 1968) showed from a study of pollen from cores at Isabella Creek at Fairbanks that a climatic change occurred about 10,000 years ago and spruce became common around 8,000 years ago. Detailed pollen work by Rampton (1971) at the Alaskan-Canadian border at the head of the Tanana River indicated a climatic amelioration around 10,000 years ago, with the invasion of shrub tundra and the invasion of a spruce woodland by 8,700 years ago. Detailed study of the late Quaternary stratigraphy in the Fairbanks area, including occurrence of megafossil remains of spruce, leads the writer to believe that scattered stands of spruce remained in lowlands in central Alaska during Wisconsinan time. Reinvasion of the forest began about 10,000 years ago (Péwé, 1975) from these lowlands. This view contrasts with Hopkins' (1972) opinion that invasion occurred around 8,500 years ago from a refugium on the exposed continental shelf (Beringia) in the vicinity of Nunivak Island.

As many as three spruce forest layers are superimposed upon one another (Giddings, 1938; Péwé, 1952a) in this Holocene formation at Fairbanks. The trees are typical of those today; the wood, which is excellently preserved in the frozen ground, differs only in color from trees of the modern interior forest. The superimposed forest layers have been cross-dated by tree-ring counting (Giddings, 1938), but the chronology outlined has not been connected with a chronology extending to present time.

Isolated pollen analyses on post-Wisconsinan sediments from localities along the highways in central Alaska indicate spruce dominance throughout the sections (Hansen, 1953; Gerard, 1954). In addition to terrestrial flora, diatoms were recovered from pond sediments of Wisconsinan age in central Alaska (Taber, 1943; Péwé, 1952a), Kenai Peninsula (Plafker, 1956), and near Palmer (Moxham and Eckhart, 1956). No stratigraphic, climatologic, or taxonomic importance can be ascribed to them at this time.

Much work remains to be done toward refining the sequences of floras in the Pleistocene of Alaska. The large nonglaciated areas will undoubtedly provide most of this material, as they have to date. There are already clear indications of changes in the position of tree line and in the composition of floras which can be related to the climate during glacial and interglacial times and the Wisconsinan thermal maximum.

Several other disciplines, in addition to palynology and the study of megafossils, may yield interesting results. Among these is the possibility of establishing a tree-ring chronology that can tie together Holocene and late Wisconsinan events. Much work remains to be done on lake sediment cores, correlating them with other areas and fitting them to a radiocarbon date chronology. With the continuing attemps to define the climatic requirements of modern arboreal and tundra species, the rather fragmentary record of vegetation distribution in Quaternary time will take on greater significance.

FAUNA

A rich fauna of Quaternary age is present in Alaska, but only in the study of the marine invertebrates and terrestrial vertebrates has a serious beginning been made. These preliminary studies have permitted statements concerning distribution and age of the fauna, and the marine invertebrates have also been instrumental in defining the times of emergence of the Bering land bridge, important to the coming of man to North America.

INVERTEBRATES

NONMARINE MOLLUSKS

Fossil land and freshwater mollusks were reported near the turn of the century (Dall, 1905), but only recently has any systematic collecting relating to stratigraphy been done. Even though little can be done at present with nonmarine mollusks in Alaskan Quaternary stratigraphy, the writer has listed and summarized all finds for ready reference.

TERRESTRIAL MOLLUSKS

Fossil land mollusks are not abundant anywhere in Alaska except in loess of post-Wisconsinan age near Gakona and Chitina in the Copper River Basin. Here the Holocene loess is peppered with white shells, but this fauna has not been studied. A few mollusks were reported from loess of probable Wisconsinan age near Tofty (Repenning and others, 1964). The writer (Péwé, 1955, p. 714) collected *Succinea avara* Say, *Discus cronkhitei* Newcomb, and *Euconulus fulvus alaskensis* Pilsbry. The last of these forms is identified by T. C. Yen of the U.S. National Museum, but the *Succinea* form is reported by Taylor (McCulloch and others, 1965, p. 450) to be less confidently identified. These species were found in loess less than 1,000 years old along the lower Delta River 160 km southeast of Fairbanks. These species also occur in loess of Wisconsinan age in Kansas (Leonard, 1952, p. 8) and Illinois (Leonard and Frye, 1960).

Fossil pulmonate snails were collected on the Seward Peninsula (D. M. Hopkins, unpub. data, 1955) and from near Rampart and Tanana (Mertie, 1937, p. 192), and *Succinea strigata* Pfeiffer (identified by J. P. E. Morrison, U.S. National Museum) was collected from loess at depths of 1–2.5 m near the Yukon River at the west end of Yukon Flats (J. R. Williams, unpub. data, 1960). Late Wisconsinan and Holocene forms were found in sediments of Kotzebue Sound (table 9) (McCulloch and others, 1965).

Until more is known about both modern and fossil forms and more collections are available, however, it is not possible to make any positive statements about the importance of Quaternary terrestrial mollusks in Alaska.

FRESHWATER MOLLUSKS

Near Fairbanks (fig. 20), retransported silt of Wisconsinan age rich in organic remains yielded the gastropods *Helisoma subcrenatum* (Carpenter), *Stagnicola palustris* (Muller), *Gyraulus parvus* (Say), and *Valvata lewisi* (Currier); the pelecypods *Sphaerium tenue* (Prime) and *Pisidium casertnum* are also present (Péwé, 1952a, p. 119–120). Taber (1943, p. 1491) also recorded the first three of these snail species in silt near Fairbanks. Mertie (1937, p. 192–193) listed fossil mol-

TABLE 9.—*Fossil nonmarine mollusks from the Kotzebue Sound area, northwestern Alaska*

[After McCulloch and others (1965, table 1). Numbers in column headings are U.S. Geological Survey Cenozoic locality numbers]

Fossil	Holocene					Late Wisconsinan to early Holocene									Wisconsinan					Sangamon					Late Illinoian			
	23165	23180	23181	23182	Summary	23125	23163	23169	23170	23171	23172	23179	23186	Summary	23162	23173	23174	23176	Summary	23126	23127	23164	23177	Summary	23178	23166	23167	Summary
Pelecypods:																												
Anodonta beringiana Middendorff	-	-	-	-	-	-	-	-	-	-	×	×	-	×	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Sphaerium lacustre (Müller)?	-	-	-	-	-	-	-	×	-	-	-	-	-	×	-	-	-	-	-	-	-	-	-	-	-	-	-	-
S. nitidum Clessin	-	-	-	-	-	-	-	×	×	-	×	-	-	×	×	-	×	×	×	-	-	×	-	×	×	×	×	×
Pisidium casertanum (Poli)	-	-	-	-	-	-	×	-	-	-	×	-	×	×	-	×	-	-	×	-	-	-	-	-	-	×	-	×
P. idahoense Roper	-	-	-	-	-	×	-	×	-	-	×	-	-	×	-	-	×	×	×	×	×	×	×	×	×	×	×	×
P. lilljeborgi Clessin	-	-	-	-	-	-	×	-	-	-	-	-	-	×	×	-	×	×	×	×	-	×	-	×	×	-	×	×
P. milium Held	-	-	-	-	-	-	×	-	-	-	×	-	×	×	-	-	-	-	-	-	-	-	-	-	-	-	-	-
P. nitidum Jenyns	-	-	-	-	-	-	-	-	-	-	×	-	×	×	-	-	×	×	×	-	×	-	-	×	-	-	-	-
P. obtusale (Lamarck)	-	-	-	-	-	-	-	-	-	-	×	-	-	×	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Freshwater gastropods:																												
Valvata lewisi Currier	×	-	-	-	×	×	×	×	×	-	×	-	×	×	×	×	×	×	×	×	×	×	×	×	×	×	-	×
Lymnaea cf. *L. turricula* Held	-	-	-	-	-	×	×	×	-	-	-	-	-	×	-	-	-	-	-	-	-	-	-	-	-	-	-	-
L. randolphi Baker	-	-	-	-	-	-	×	-	-	×	×	-	?	×	-	-	×	-	×	×	×	-	-	×	×	-	?	×
L. stagnalis (Linnaeus)	-	-	-	-	-	-	-	×	-	-	-	-	-	×	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Gyraulus circumstriatus (Tryon)	-	-	-	-	-	-	-	-	-	-	-	-	×	×	-	-	-	-	-	-	-	×	-	×	-	-	-	-
G. parvus (Say)	-	-	-	-	-	×	×	?	×	-	×	-	-	×	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Armiger crista (Linnaeus)	-	-	-	-	-	-	-	×	×	-	×	-	×	×	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Helisoma subcrenatum (Carpenter)	-	-	-	-	-	-	-	×	×	-	-	-	×	×	-	-	-	-	-	-	-	-	-	-	-	×	-	×
Physa skinneri Taylor	-	-	-	-	-	-	×	-	-	-	-	-	-	×	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Land gastropods:																												
Discus cronkhitei (Newcomb)	×	-	-	-	×	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
cf. *succinea*	×	×	×	×	×	-	-	-	-	×	×	-	-	×	-	-	-	-	-	-	-	-	-	-	-	-	-	-

lusks collected by P. S. Smith from Quaternary silt deposits in this area.

Gyraulus parvus (Say) and *Valvata lewisi* (Currier) were found in lake sediments radiocarbon dated as 8,000 years old near Matanuska glacier (J. R. Williams, written commun., 1964). F. R. Weber reported the freshwater pelecypod *Pisidium lilljeborgii* Clessin from late Quaternary sediments near the Porcupine River and from Cave–off Cliffs of the lower Yukon (oral commun., Jan. 20, 1963). Weber also collected *Valvata lewisi* (Currier) from late Quaternary sediments on the Porcupine River and at two localities in the Yukon-Koyukuk lowland. Hopkins (unpub. data, 1957) reported *Pisidium* sp., *Valvata lewisi* (Currier), *Valvata helicoidea,* and *Stagnicola palustris* (Muller) from Quaternary sediments near Tofty.[4]

A large collection of pelecypods and gastropods, made by McCulloch, Taylor, and Rubin (1965) near Kotzebue Sound, represents the first data available on specimens of pre-Wisconsinan age. As they point out (p. 442), however, our knowledge of geographic and stratigraphic differences between fossil and living groups in Alaska is so incomplete that we cannot at present use the faunas to interpret changes of climate and changes in geographic range of species.

Sediments of Wisconsinan and post-Wisconsinan age from the many lakes in Alaska are beginning to yield data on their fossil faunas (Livingstone and others, 1958; Charlotte Homquist, unpub. data, 1959).

Very little is known of the modern distribution of freshwater mollusks in Alaska, and this information seems essential to the interpretation of the fossil record. The gastropods *Stagnicola palustris* (Muller) and *Helisoma subcreatum* (Carpenter) live today in cold-water ponds in central Alaska, near Fairbanks and Galena (pl. 1). Hopkins (1963) stated that species of *Valvata* and *Pisidium* live in lakes in central Seward Peninsula today. In 1948 the writer collected specimens of *Anodonta beringiana* Middendorf from a clear-water tributary of the Yukon 16 miles southwest of Beaver, which represents one of the few localities in Alaska for a species that is probably widespread in both Alaska and Siberia (R. D. Reger, oral commun., Dec. 10, 1969). It is thought that this species migrated across the land bridge to Alaska in Illinoian or Wisconsinan time; these mollusks have not migrated farther east than the approximate position of the Wisconsinan continental ice sheet near the Canada-Alaska border. R. D. Reger and D. W. Taylor have compiled a wealth of unpublished information on freshwater mollusk distribution; this type of study has great promise for Alaska.

The only study of living terrestrial and freshwater mollusks associated with modern glacier termini is the recent work of Tuthill (1970) near Martin Glacier in southern Alaska. This study was very helpful in relat-

[4]Mollusks collected by Weber were identified by D. W. Taylor; those by Hopkins were identified by Taylor and by G. D. Hanna. The writer's specimens were identified by J. P. E. Morrison and R. D. Reger.

ing conditions of present glacier fronts to those that existed in Pleistocene time.

MARINE MOLLUSKS

The record of the Quaternary marine mollusks is much more completely known and useful than that of the terrestrial and freshwater mollusks. Marine ostracodes are reported but not yet widely used in stratigraphy (Swain, 1961; Schmidt, 1963; Schmidt and Sellmann, 1966). The foraminiferal fauna of the marine deposits of Quaternary age from northern Alaska was studied (Faas, 1962) and can be summarized as differing little, if at all, from those living in the same area today. This is equally true of the foraminiferal fauna from glaciomarine sediments from Juneau (Smith, 1970). Quaternary marine mollusks have been reported in Alaska since the turn of the century, but these data, plus later studies, were not of maximum stratigraphic use until the collections were reevaluated, compared with more recent collections, and put into stratigraphic position (Hopkins, 1959b; Hopkins and others 1960; Hopkins, 1965, 1967a).

The record of the marine mollusks of Quaternary age in Alaska is from deposits formed by transgressions of the sea (p. 75). The information presented here is summarized from reports by Hopkins (1959b), Hopkins, MacNeil, and Leopold (1960), Hopkins (1965, 1967a), and McCulloch (1967).

Marine invertebrates in deposits of the Beringian transgression (late Pliocene) have a decidedly modern aspect, but most have subtle differences from the living species. Some of the species are extinct. The stratigraphic distribution of the more significant fossils in marine sediments of Quaternary age in western and northern Alaska was summarized by Hopkins (1967a, table 2), Hopkins, MacNeil, and Leopold (1960, table 1), and McCulloch (1967, table 1). (See table 10.)

Many of the extinct forms are unique to the deposits of the Beringian age and to Alaska. Some are related to Pliocene species of Japan or to the Pliocene and Pleistocene of California. Several living forms that are common in younger beds make their first appearance in Alaska in the deposits of Beringian age.

Perhaps the most interesting observation to be made concerning the fauna of this age is in connection with the Bering land bridge as a barrier to migration of marine organisms. Hopkins (1959b, 1967b) pointed out that the bridge served as a barrier for marine organisms in the Pacific and Arctic Oceans. The presence of the large scallop *Patinopecten (Fortipecten) hallae* (Dall) in Beringian deposits at Solomon and Kivalina north of Kotzebue indicates that by late Pliocene time a new seaway extended across the Bering-Chukchi platform, permitting marine organisms that had previously been

TABLE 10.—*Stratigraphic distribution of the more significant fossils found in marine beds of Pleistocene age in Alaska* [*From Hopkins (1967b)*]

Column headings: (1) Beringian, (2) Anvilian, (3) Einahnuhtan, (4) Kotzebuan, (5) Pelukian, (6) Woronzofian, (7) living in Alaskan waters at present time. *Symbols:* Tapered ends of line indicate first or last recorded appearance in Alaska; solid lines indicate presence of animal firmly established; queried lines indicate questionable identifications or questionable age assignment for enclosing beds; dashed lines indicate animal probably present but has not been collected. *Notes:* E—animal extinct; EC—animal extinct but closely related to a living form; A—nearest relative now confined to North Atlantic; J—nearest relative now confined to Japanese or Eastern Siberian waters.

Species	1	2	3	4	5	6	7	Notes
GASTROPODA								
Margaritopsis grosvenori Dall								
Littorina sp. aff. *L. palliata* (Say) and *L. mandschurica* Schenck								A or J
Epitonium (Boreoscala) groenlandicum (Perry)								
Tachyrhyncus erosus (Couthouy)								
Trichotropis insignis Middendorf								
T. bicarinatus (Sowerby)								
Amauropsis purpurea Dall								
Natica (Cryptonatica) janthostoma Deshayes								J
Trophon kamschatkanus Dall								J
Buccinum angulosum Gray								
Buccinum fringillum Dall								
Buccinum glaciale Linne'								
Pyrulofusus schraderi Dall								E
Pyrulofusus deformis (Reeve)								
Beringius kobelti (Dall)								
Plicifusus sp. aff. *P. wakasanus* Dall								J
Neptunea leffingwelli (Dall)								J
Neptunea heros mesleri (Dall)								EC
Neptunea heros (Gray)								
Neptunea spp. [1]								E
Volutopsius sp. aff. *V. stefanssoni* Dall								EC
Volutopsius stefanssoni Dall								
Propebela sp. cf. *P. exquisita* Bartsch or *P. assimilis* (Sars)								J or A
PELECYPODA								
Yoldia koluntunensis Slodkewitsch								E
Portlandia arctica (Gray) [2]								
Chlamys (Leochlamys) tugidakensis MacNeil								E
Chlamys (Swiftopecten) kindlei (Dall)								EC, J
Chlamys ("Chlamys") coatsi MacNeil								E
Chlamys ("Chlamys") coatsi middletonensis MacNeil								E
Chlamys ("Chlamys") lioica Dall								E
Chlamys (Chlamys) beringiana colvillensis MacNeil								EC
Chlamys (Chlamys) hanashienses amchitkana MacNeil								EC
Chlamys (Chlamys) plafkeri MacNeil								E
Chlamys (Chlamys) islandica kanagae MacNeil								EC
Chlamys (Chlamys) islandica powersi MacNeil								EC
Fortipecten hallae (Dall)								E
Astarte actis Dall								E?
Astarte nortonensis MacNeil								EC
Astarte borealis Schumacher[3]								
Astarte broweri Meek								EC
Astarte leffingwelli Dall								E
Astarte hemicymata Dall								E
Astarte sp. cf. *A. soror* Dall								A
Astarte sp. aff. *A. bennetti* Dall								EC
Astarte bennetti Dall								
Protothaca adamsi (Reeve)								J
Tellina (Peronidea) lutea Gray								
Macoma sp. cf. *carlottensis* Whiteaves								
Macoma baltica (Linne')								
Siliqua patula (Dixon)								
Mya (Arenomya) japonica Jay								
Cyrtodaria kurriana (Dunker)								
ECHINODERMATA								
Echinocyamus sp. cf. *E. pusillus* (Müller)								A
CIRRIPEDIA								
Verruca alaskana Pilsbry								E

[1] Beringian and Anvilian beds contain complexes of *Neptuneas*, in addition to those listed above that are ancestral to but distinct from several modern species.

[2] *Portlandia arctica* has not been found in Alaskan beds older than the Kotzebuan transgression, but it is present in Chukotka in the Pinakul' beds of Einahnuhtan age.

[3] *Astarte borealis* is found throughout the Pleistocene marine sequence of Tjörnes, Iceland (Einarsson and others, 1967), but this stock is represented in the older part of the Alaskan Pleistocene sequence by the closely related *A. nortonensis*.

confined to the north Pacific Ocean to migrate northward.

Marine invertebrates of the Anvilian transgression are best known from Nome and the Arctic coast and consist largely of forms living in adjoining waters at the present time. Some new forms first appear here. Extinct *Neptunea heros mesleri* and *Astarte leffingwelli* occur in deposits of this age along the Chukchi Sea (McCulloch, 1967) but not in deposits younger than those of the Einahnuhtan transgression. Several fossil forms are closely related to species now limited to waters adjoining Japan, or to the North Pacific Ocean or southern Bering Sea.

The faunas of the Einahnuhtan transgression are essentially modern in character (table 10).

Marine invertebrates of the Kotzebuan transgression (Yarmouth age) are best recorded in deposits from Kotzebue Sound. A rich molluscan fauna there consists entirely of species living in nearby waters at the present time. No extinct forms are present. *Portlandia arctica* (Gray) and others made their first recorded appearance in western Alaska at this time.

Molluscan faunas of the Pelukian transgression (Sangamon age) consist entirely of living forms, with the exception of the extinct *Cardita (Cyclocardia) subcrassidens* MacNeil. Fossils of this age have been reported from northern and western Alaska and Pribilof Islands. Several forms are limited in their present distribution to areas well south of their northernmost fossil occurrence. *Mya japonica* Jay made its first Alaskan appearance in Pelukian beds.

For the most part, marine mollusks of Quaternary age in Alaska record cool-water environments. Early Quaternary and late Pliocene forms indicate a warmer marine environment, but mollusks of middle and late Quaternary time indicate water temperatures about the same as now except perhaps during Sangamon time.

The fauna is essentially modern in character, with a few to many extinct forms recorded in deposits of early Quaternary and late Pliocene age, but with only a rare extinct form in deposits of middle to late Quaternary age. An understanding of Quaternary marine deposits and associated fauna in western Alaska is now available and may stimulate study of the other coastal areas in the State.

INSECTS

A new and potentially important method of paleoenvironmental study of the Quaternary deposits of Alaska is based on examination of fossil insects, mainly beetles, from the frozen peats and silts. Well-preserved fossil insects occur in sediments in Alaska, but no studies have been done until recently.

Spurred by the success of recent studies of insects of Pleistocene age in England, work was initiated on the frozen deposits near Fairbanks. A pioneer effort by Matthews (1968a, b) has led to enlarged studies elsewhere in Alaska. Beetles, because of their rather specific habitat preferences, may be more sensitive to climatic change than plants and may eventually provide a more detailed record than pollen.

Matthews' study (1968a, b) of the carabid beetles (ground beetle) from the retransported silt of Wisconsinan age (fig. 20) from a mining exposure at Eva Creek (fig. 29) near Fairbanks indicated that there was a tundra environment during Wisconsinan time, that coniferous tree line must have been below the present elevation of the Eva Creek, and that the area was essentially treeless. This interpretation of the environment is also borne out by analysis of pollen from Eva Creek by Matthews (1968b, p. 222) and earlier work by E. S. Barghoorn (written commun., Nov. 23, 1954) on pollen samples collected by the writer. Matthews (1974) has now extended his study of insects to deposits at Cape Deceit in western Alaska and obtained challenging data on paleoenvironments of Pleistocene time. The writer believes that a great opportunity exists to extend this type of work elsewhere in the State and that only the beginning has been made in such paleoenvironmental studies.

BACTERIA IN PERMAFROST

The possibility that viable bacteria may exist in permafrost of Pleistocene age has intrigued microbiologists for many years, for such bacteria could be the oldest living organisms in the world. Investigations by Kriss (1940) in the Arctic of U.S.S.R. and James and Sutherland (1942) in Churchill, Canada, revealed bacteria in permafrost, but no age assignment is available. Becker and Volkmann (1961) recovered eight bacteria of four genera from a core taken from permafrost at the U.S. Army Cold Regions Research and Engineering Field Laboratory on Farmer's Loop Road at Fairbanks. Becker continued the project for several years with drilling at Ready Bullion Creek (Péwé, 1965a, p. 24) near Fairbanks but felt the absence of true sterile conditions prevented a positive proof of existence of viable bacteria in permafrost (oral commun., 1965). Boyd and Boyd (1964) carefully examined permafrost in large drill holes near Barrow and found a few living bacteria but noted that these may have been from the surface and not of Pleistocene age. Kjøller and Ødum (1971) carefully sampled permafrost from Ready Bullion Creek and believe it probable that living microorganisms are present in the frozen Pleistocene sediments. Obviously, too little is known of this subject and more detailed work should be undertaken.

VERTEBRATES

Alaska, like northern Siberia, has long been famous for the abundant remains of extinct Pleistocene mammals, found in frozen deposits along major rivers and in the valleys of many minor streams. The earliest account of these fossils seems to be that of Kotzebue (1821, p. 218–220), who found abundant vertebrate remains at Elephant Point in Eschscholtz Bay during his expedition to the Chukchi Sea in 1816. F. W. Beechey also collected there, and the mammal bones were reported by Buckland (1831). Because early explorers (Dall, 1869, 1873) reported a great abundance of bones, several expeditions were conducted in Alaska in the hope of finding complete skeletons or perhaps even frozen carcasses comparable with those found in Siberia (Madren, 1905; Gilmore, 1908; Quackenbush, 1909). When large-scale gold mining began in the Fairbanks district in 1928, extensive fossil collecting was undertaken there and elsewhere in Alaska by the late O. W. Geist and others on behalf of the American Museum of Natural History. A small part of this material has been described by Frick (1937) and Skinner and Kaisen (1947). Later, Geist and others collected vertebrate fossils from the Fairbanks area and northern Alaska for the Museum of the University of Alaska. Some of the most detailed work ever done on the late Pleistocene mammals in central and western Alaska was that by Guthrie (1966a, b, c; 1967; 1968a, b; Guthrie and Matthews, 1971), who studied collections that he had made, as well as the vast collection of Geist.

DISTRIBUTION

Most of the remains of land mammals are from the unglaciated part of Alaska (fig. 42). Such distribution is to be expected because animals were mostly absent in glaciated areas during glacial maximums and glacial advances tended to destroy or cover earlier fossil remains.

The greatest collection of vertebrate specimens is from the Fairbanks area, where tens of thousands of specimens have been collected during the past 30 years. For example, in 1938, a typical year, 8,008 cataloged specimens weighing about 8 tons were collected by O. W. Geist and shipped to the American Museum of Natural History in New York City (University of Alaska, "Collegian," 1938, fall). Partial lists of mammals from the Fairbanks area were given by Frick (1930, 1937), Wilkerson (1932, p. 422), Mertie (1937, p. 191), Stock (1942), Hibben (1943), Taber (1943, p. 1487), Skinner and Kaisen (1947), Skarland (1949, p. 132–133), Péwé (1952a, table 4), Geist (1953), Péwé and Hopkins (1967), and Guthrie (1968a).

The geological literature (U.S. Geological Survey Bulletins) in Alaska dealing with early placer mining activities mentions in passing that bones of extinct animals such as mammoth, mastodon, bison, horse, and others were found in many localities in addition to the Fairbanks area. Vertebrate remains on the Seward Peninsula were mentioned by Collier (1902), Quackenbush (1909), Hopkins (1963), Harington (1970a, b), and many others. Mertie (1937, p. 190–191) summarized finds of several specimens from the Yukon-Tanana Upland. Whitmore and Foster (1967) listed finds from Chicken and Lost Chicken Creek, including *Panthera atrox,* and Repenning, Hopkins, and Rubin (1964) listed fossils from Tofty.

Chapman and Sable (1960, p. 124) mentioned mammoth tusks found along banks of the Utukok and Kokolik Rivers of northern Alaska. Other references to vertebrates on the Arctic Coastal Plain (pl. 1) were made by Livingstone, Bryan, and Leahy (1958), Harington (1969), and others. William Quaide (unpub. data, 1953) collected mammal remains along the Kuk and Ikpikpuk Rivers in northern Alaska, and Robert Fladeland identified wolf, bear, large cat, mammoth, horse, moose, caribou, musk ox, and bison. A *Saiga* bone was also identified in this collection by C. R. Harington (C. E. Ray, oral commun., Feb. 12, 1970). C. A. Repenning (written commun. to D. M. Hopkins, Feb. 5, 1962) provisionally identified many vertebrate remains collected by W. J. Maher from the Ikpikpuk River including *Smilodon* sp. and *Felis (Lynx) lynx.* Ray (1971) reported that bones of the woolly mammoth were found on St. Lawrence, Pribilof, and Unalaska Islands on the Bering-Chukchi platform.

STRATIGRAPHY

Unfortunately, the stratigraphic context of most of the vertebrate material of Pleistocene age from Alaska now in museums is unknown. Fossils are most readily seen when the enclosing matrix has been washed away, and so most of the vertebrate fossils have been found lying loose on beaches or riverbanks and the floors of placer-mining excavations. Furthermore, detailed studies of Alaskan Quaternary stratigraphy did not begin until the late 1940's; thus even when bones were collected in situ in earlier years, no conclusions could be drawn as to the age of the enclosing matrix. Skinner and Kaisen's (1947) thorough study of fossil *Bison* is based upon Alaskan material whose stratigraphic context is not known; their speculations on *Bison* evolution are therefore based entirely upon the morphology of the skeletal material and upon assumptions as to probable evolutionary trends in critical dimensions such as horn width and tooth size. The great collection of Pleistocene mammal bones from the Fairbanks area in the Frick Laboratory of the American Museum of Natural History will be important in many taxonomic studies, but stratigraphic relations are missing.

During the last two decades, a modest but growing number of vertebrate fossils have been found in significant stratigraphic contexts, most in the Fairbanks area but a few elsewhere. Most were in beds of Wisconsinan and Holocene age, but a significant few were in beds of Sangamon, Illinoian, and pre-Illinoian age (Péwé and Hopkins, 1967; Guthrie and Matthews, 1971). Extensive collections of bones of small mammals from silts near Fairbanks by Guthrie (1968b) are from sections where stratigraphy has been established. (Péwé, 1952a).

PRE-ILLINOIAN

Vertebrate remains of pre-Illinoian age are rare. Hopkins (in Péwé and Hopkins, 1967, p. 268–269) found bones of *Bison* (*Superbison*) sp., *Rangifer*, and *Dicrostonyx torquatus* in beds of the Kotzebuan transgression at Kotzebue Sound, which he interpreted to have been deposited during a pre-Illinoian(?) interglaciation.

Mammoth bones were found in the perennially frozen auriferous gravel in the Fairbanks area by early workers (Prindle, 1908, p. 32). The writer collected specimens of mammoth and bison from the gravel deposits that underlie Illinoian and Yarmouth(?) beds and thinks that these gravel units represent sediments laid down during one or more pre-Illinoian glacial intervals. These finds of *Bison* in central and western Alaska are perhaps the earliest recorded evidence of these genera in North America. The finds of *Rangifer* and *Dicrostonyx* were until recently the earliest recorded in North America.

The only record of early Pleistocene mammals in Alaska was reported by Guthrie and Matthews (1971) from Cape Deceit on the south shore of Kotzebue Sound near Deering (pl. 1). A complex sequence of peat and organic rebedded loess (the Cape Deceit Formation) contains representatives of the genera *Canis, Rangifer, Cervus, Ochotona, Lemmus, Microtus,* and *Pliomys* and a newly described genus and species, *Predicrostonyx hopkinsi*—predecessor of the extant genus *Dicrostonyx* (table 11). Guthrie and Matthews stated that, in addition, new species of *Ochotona, Pliomys,* and *Microtus* occur. From the stage of evolution of the fossils and from a study of the stratigraphy, it is assumed that the Cape Deceit Formation is at least pre-Cromerian in age. The Cape Deceit fauna contains the earliest North American record of *Rangifer, Equus,* and *Microtus* and the only North American record of *Pliomys*.

ILLINOIAN

The list of documented finds of fossil mammal remains of Illinoian age is slowly growing. Péwé (1952a, 1958a, 1965a) reported such occurrences, and with continued work on Pleistocene mammal finds and interest in land bridge migrations (Hopkins, 1967a), a clear picture is beginning to appear. Much more work, however, is necessary to understand better the Palearctic and Neararctic mammalian dispersal in late Cenozoic time.

All the finds of mammalian fossils known to the writer to be from Alaska sediments recognized as pre-Wisconsinan are given in table 11. The evidence upon which the stratigraphic dating is based was discussed by Péwé (1952a, 1965a; Fairbanks area), Wahrhaftig (1958; Nenana Valley), Williams (1962; Chandalar Valley), Hopkins, MacNeil, and Leopold (1960; Nome area), McCulloch, Taylor, and Rubin (1965) and Guthrie and Matthews (1971; Kotzebue Sound area), and Péwé, Hopkins, and Giddings (1965; Alaska in general). Many of the finds were discussed by Péwé and Hopkins (1967).

From table 11, it is evident that many taxa were present in North America earlier than is generally recognized. One of the most recent good summaries of Quaternary mammals in North America (Hibbard and others, 1965, tables 2, 3; Flint, 1971, p. 766–768) indicated no pre-Wisconsinan record for *Rangifer, Ovibos, Ovis, Alces, Saiga, Bootherium, Symbos,* and *Dicrostonyx* and no pre-Illinoian record for *Bison*. Kurtén (1968, p. 268–269) also stated that *Bison* did not migrate to North America until Riss time and that *Rangifer, Ovibos,* and *Saiga* did not migrate to Alaska from Eurasia until Würm time. However, all these genera listed by Hibbard and Kurtén were present in Alaska by at least Illinoian time and represent Eurasian immigrants. *Rangifer* was in North America in early Pleistocene time. The find of *Bison* in the Fairbanks area in deposits of pre-Illinoian age is consistent with Flerow's (1967) idea that *Bison priscus crassicornis* migrated from Asia to America before the Riss Glaciation.

The Siberian steppe antelope, *Saiga* (fig. 42), was found in the Pleistocene deposits of Alaska. The earliest published find was that of O. C. Kaisen and O. W. Geist at Lillian Creek near Fairbanks in 1930 (described in Frick, 1937). Geist and others also collected *Saiga* from Gold Hill in 1952, Cripple Creek Sump in 1949, Gilmore Creek in 1931, and Goldstream Creek in 1933 (M. F. Skinner, written commun., Mar. 3, 1970). Recently, Guthrie (1968a, p. 353) identified bones of *Saiga* collected by Geist from Cripple Creek in Gold Hill. John Dorsh collected the left metatarsus of *Saiga* from Banner Creek in 1935. *Saiga* was identified in the collection by William Quaide from the Arctic Coastal Plain from Alaska (C. E. Ray, oral commun., Feb. 12, 1970). The writer believes the Cripple Creek Sump and Gold Hill specimens are Illinoian in age; other specimens could be either Illinoian or Wisconsinan in age, most likely the latter. C. R. Harington of the National Museum of Canada is describing the Alaska specimens (written commun., July 10, 1970).

Saiga is found nowhere in North America but Alaska; this area represents the easternmost extension of its Siberian range (Sher, 1967, p. 110). Southward penetration from Alaska by this animal was probably prevented by mountains, thick snow blankets, glaciers, or forests, because *Saiga* requires an environment of

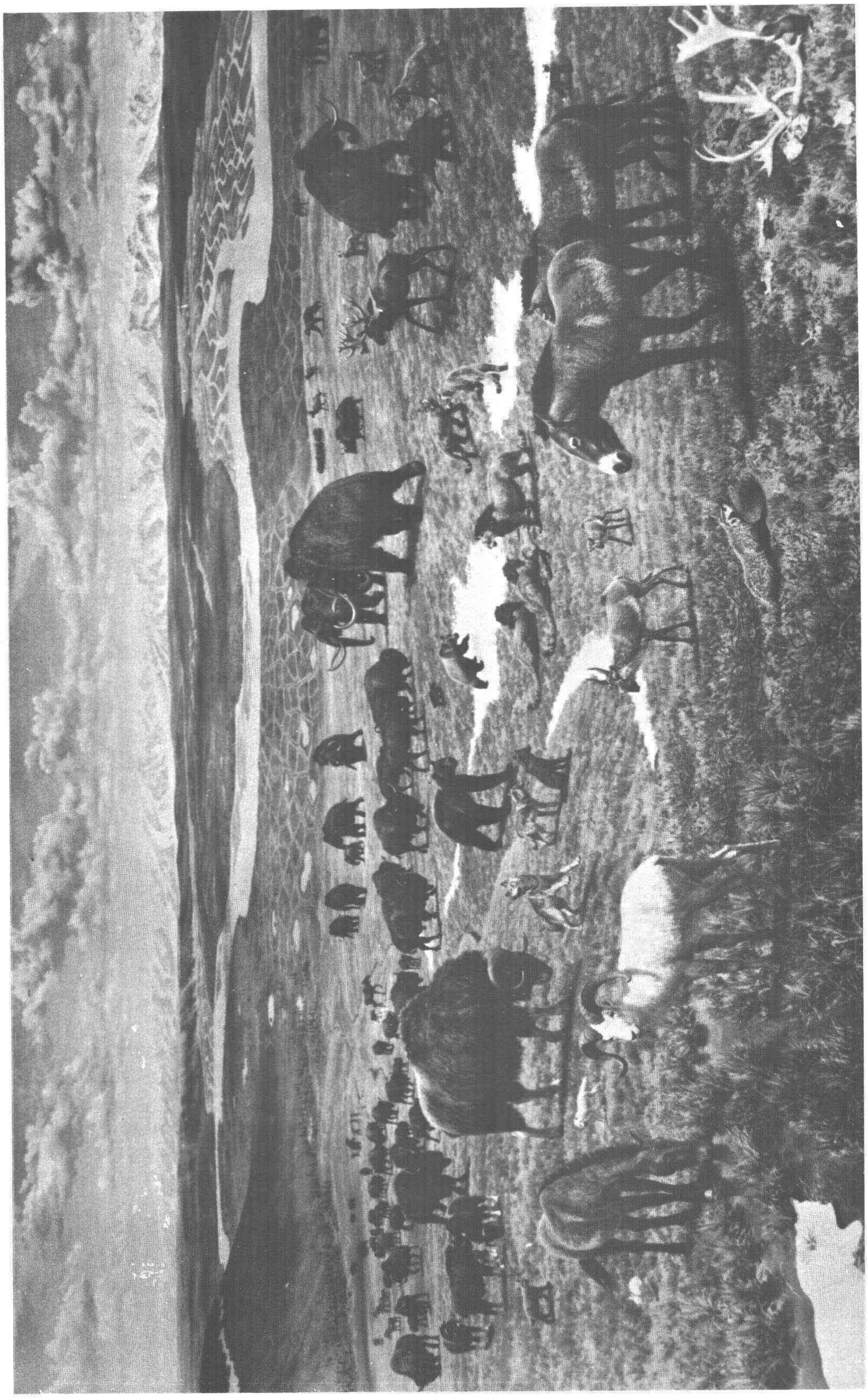
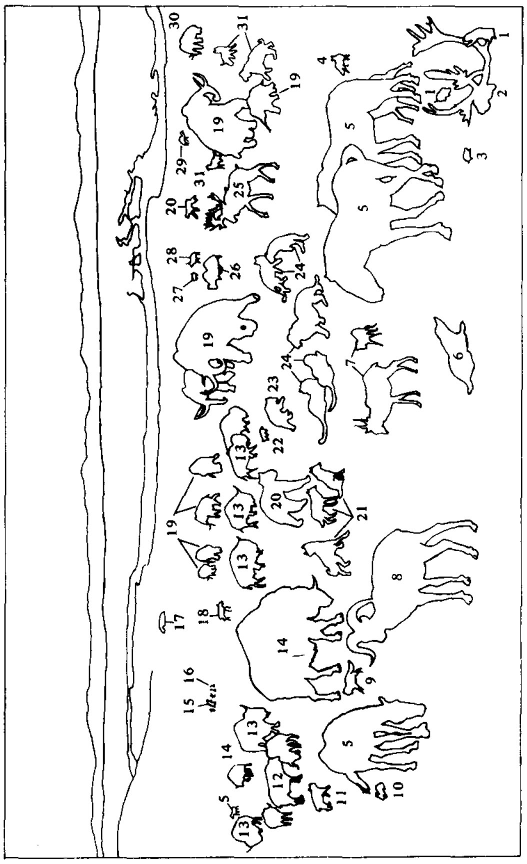


FIGURE 42.—Mammals of late Wisconsinan time. View is toward the southeast, across the westward flowing Tanana River near Fairbanks; Alaska Range is in background. Mural by J. H. Matternes. (Reproduced through the courtesy of National Geographic Society, copyright 1972.)



Key to animals shown:

● *Extinct*

▲ *No longer native to Alaska*

1. Arctic ground squirrel, *Citellus undulatus*
2. Antlers of caribou, *Rangifer tarandus*
3. Brown lemming, *Lemmus sibiricus*
4. Red fox, *Vulpes vulpes*
5. Horse, *Equus* ▲
6. Badger, *Taxidea taxus* ▲
7. Saiga antelope, *Saiga ricei* ▲
8. Dall sheep, *Ovis dalli*
9. Arctic fox, *Alopex lagopus*
10. Alaskan tundra hare, *Lepus othus*
11. Lynx, *Lynx canadensis*
12. Musk ox, *Bootherium nivicolens* ●
13. Large-horned bison, *Bison priscus* ●
14. Musk ox, *Symbos cavifrons* ●
15. Ground sloth, *Megalonyx* ●
16. Man, *Homo sapiens*
17. Beaver, *Castor* (only lodge shown)
18. Moose, *Alces alces*
19. Woolly mammoth, *Mammuthus primigenius* ●
20. Great North American short-faced bear, *Arctodus simus* ●
21. Wolf, *Canis lupus*
22. Wolverine, *Gulo gulo*
23. Grizzly bear, *Ursus arctos*
24. Lionlike cat, *Panthera atrox* ●
25. Stag-moose, *Cervalces alaskensis* ●
26. Yak, *Bos bunnelli* ▲
27. Musk ox, *Ovibos moschatus* ▲
28. Wapiti, *Cervus elaphus* ▲
29. Camel, *Camelops* ●
30. American mastodon, *Mammut americanum* ●
31. Saber-toothed cat, *Homotherium serum* ●

Courtesy of the National Geographic Society

plainlike relief with a hard surface, little forest, and little snow. The writer believes that such an environment existed in late Pleistocene time in Alaska on the windswept, flat, braided flood plains of glacial streams that were abundant at that time.

The yaklike *Bos* is another Siberian animal that was rare in Alaska in Pleistocene time and that did not move southward into North America. Since Frick's early description, other bones in the Geist collection have been identified as *Bos* from four different creeks in the Fairbanks area (Guthrie, 1968a), but they have no stratigraphic context; remains from Gold Hill perhaps are Illinoian. Future work on the Alaskan collection in the American Museum of Natural History may enlighten our concept of *Bos* in Alaska.

Perhaps the only Pleistocene mammal that may have been restricted to pre-Wisconsinan time in Alaska is the extinct musk ox, *Praeovibos*. C. E. Ray (written commun., Oct. 25, 1967) reported *Praeovibos* was quite rare, for all the identified specimens are from the Fairbanks area; only five brain cases or frontlets can be identified with reasonable certainty, and two fragmentary specimens are doubtfully referable to this genus. The five specimens come from Cripple Creek Sump, one example from Gold Hill, and are probably Illinoian in age. A specimen from lower Cleary Creek may be either Illinoian or Wisconsinan, but no stratigraphic information is available.

Until recently, *Praeovibos* was known from only the middle Pleistocene deposits of Western Europe (Kahlke, 1963), but Sher since found *Praeovibos* in middle and eastern Siberia in deposits of pre-Illinoian age (Vangengeim and Sher, 1970; Sher, 1969, 1971).

WISCONSINAN

DISTRIBUTION

Most of the fossil vertebrate remains of Quaternary age in Alaska occur in frozen silt deposits of Wisconsinan age and are common in central, western, and northern Alaska. From Seward Peninsula Hopkins (1963) reported mammoth, horse, and bison remains in deposits of known Wisconsinan age.

The most abundant large mammal remains in the Fairbanks area are those of the bison (fig. 42). The mammoth (*Mammuthus primigenius*) and horse (*Equus alaskae* and other species) are next in abundance (Péwé, 1952a, p. 121). *Bison (Superbison) crassicornis* is the most abundant of the four bison types defined by Skinner and Kaisen (1947, p. 141), making up 87 percent of the total bison skulls. All four bison subgenera established by Skinner and Kaisen from central Alaska are known to occur in sediments of Wisconsinan age. The revision of the bison classification by Skinner and Kaisen is not accepted by all workers; Guthrie (1966a,

TABLE 11.—*Mammal remains found in beds of pre-Wisconsinan age in Alaska*

[From Péwé and Hopkins (1967), with additions (footnoted)]

Fossil	Age	Location
Order Rodentia:		
Castor sp (beaver)	Illinoian	Fairbanks area.
Marmota sp. (marmot)	Early Pleistocene	Cape Deceit.[1]
Ondrata zibethecus (muskrat)	Late Illinoian or Sangamon	Kotzebue Sound area.
Pliomys sp. (meadow mouse)	Middle Pleistocene	Cape Deceit.[1]
deeringensis (meadow mouse)	Early Pleistocene	Do.
Microtus gregalis (vole)	Illinoian	Fairbanks area.[2]
sp. (vole)	Middle Pleistocene	Cape Deceit.[1]
deceitensis (vole)	Early Pleistocene	Do.
Lemmus sibericus (lemming)	Illinoian	Fairbanks area.[2]
cf. *sibericus* (lemming)	Early Pleistocene	Cape Deceit.[1]
Dicrostonyx torquatus (collared lemming)	Pre-Illinoian(?) interglaciation	Kotzebue Sound area.
Do	Illinoian	Fairbanks area.[2]
Do	do	Cape Deceit.[1]
Discrostonyx henseli	do	Do.
Predicrostonyx hopkinsi	Early Pleistocene	Do.
Citellus undulatus (ground squirrel)	Illinoian	Fairbanks area.[2]
Do	do	Fairbanks area.[3]
sp. (ground squirrel)	Early Pleistocene	Cape Deceit.[1]
Order Lagomorpha:		
Ochotona whartoni (pika)	do	Do.
Do	Illinoian	Fairbanks area.[1]
Order Insectivora:		
Sorex (Sorex) sp. (shrew)	Early Pleistocene	Cape Deceit.[1]
Order Carnivora:		
Ursus sp. (bear)	Illinoian	Fairbanks area.[4]
Xenocyon sp. (hunting dog)	do	Fairbanks area.
Canis sp. (wolf)	do	Fairbanks area.[5]
Canis(?) sp. (wolf)	Early Pleistocene	Cape Deceit.[1]
Vulpes sp. (fox)	Illinoian	Fairbanks area.
Felis sp. (large cat)	do	Fairbanks area.[5]
Order Proboscidea:		
Mammut sp. (mastodon)	do	Fairbanks area.
Mammuthus sp. (mammoth)	do	Nenana valley.
Do	do	Fairbanks area.[5]
Do	Pre-Illinoian glaciation	Fairbanks area.
Order Artiodactyla:		
Cervus sp. (elk)	Illinoian	Fairbanks area.[5]
cf. *elaphus* (elk)	Early Pleistocene	Cape Deceit.[1]
Cervalces sp. (giant elk)	Illinoian	Fairbanks area.
Alces sp. (moose)	do	Do.
Rangifer sp. (caribou)	do	Fairbanks area.[5]
Do	Pre-Illinoian(?) interglaciation	Kotzebue Sound area.
Do	Early Pleistocene	Cape Deceit.[1]
Saiga sp. (steppe antelope)	Illinoian	Fairbanks area.[4]
Bison (Superbison) sp. (large bison)	Sangamon	Tofty mining district.
Do	Illinoian	Fairbanks area.[5]
Do	Pre-Illinoian glaciation	Fairbanks area.
Do	Pre-Illinoian(?) interglaciation	Kotzebue Sound area.
Bootherium sp. (extinct musk ox)	Illinoian	Fairbanks area.
Ovibos moschatus (modern musk ox)	do	Nome area.[6]
sp. (musk ox)	do	Fairbanks area.[5]
Symbos sp. (musk ox Bovid)	do	Fairbanks area.[7]
Praeovibos sp. (extinct musk ox)	do	Do.
Ovis sp. (mountain sheep)	do	Fairbanks area.
cf. *Ovis* sp. (mountain sheep)	do	Nenana valley.
Order Perissodactyla:		
Equus lambei (horse)	Illinoian(?)	Yukon Flats.
sp. (horse)	Illinoian	Fairbanks area.[5]
Do	Early Pleistocene	Cape Deceit.[1]
Order Pinnipedia:		
Odobenus sp. (walrus)	Sangamon	Nome.[8]
Eumetopias sp. (Steller's sea lion)	do	St. Paul Island.[9]
Order Sirenia:		
Hydrodamalis sp. (Steller's sea cow)	do	Amchitka Island.[10]

[1]Guthrie and Mathews (1971).
[2]Guthrie (1968b, p. 232).
[3]Péwé, Fairbanks Creek, Fairbanks area, June 10, 1952.
[4]O. W. Geist (written commun., Dec. 5, 1950). Stratigraphic interpretations by Péwé. Samples from Cripple Creek Sump.
[5]In addition to specimens from Cripple Creek Sump given in Péwé and Hopkins (1967), Péwé and Geist collected bison, horse, mammoth, lion, fox, muskox, caribou, wolf, and elk from loess of Illinoian age in Gold Hill mining cut (unpub. data, June 1, 1952).
[6]Harington (1970b).
[7]C. E. Ray (written commun., Oct. 25, 1967). Stratigraphic interpretations by T. L. Péwé. Samples from Cripple Creek Sump.
[8]D. W. McCulloch (written commun., 1967).
[9]D. M. Hopkins (written commun. from C. A. Repenning, Apr. 20, 1970).
[10]Gard, Lewis, and Whitmore (1972).

p. 739) believed that all the large extinct bison fossils in Alaska represent one species, *Bison priscus* (=*B. crassicornis*).

Except for *Citellus*, the remains of small mammals had not been described in stratigraphic context from the Fairbanks area. Guthrie (1968b) made an exhaustive study of rodents that he carefully collected from known stratigraphic horizons. He found the following in sediments of Wisconsinan age: *Microtus gregalis* (predominant member of the assemblage), *Lemmus sibericus, Dicrostonyx torquatus, Citellus undulatus.*[5]

[5]Various species and subspecies names have been used for the ground squirrels in Alaska (Rausch, 1953). *Citellus parryi* (Richardson) was used in the older literature (Hill, 1942) but *C. undulatus* in the newer (Guthrie, 1968b). Rausch (1953, p. 121) pointed out that *C. undulatus* has priority over *C. parryi*. However, Liapunova and Vorontsov (1970), studying chromosomes of the *Citellus* and evolution of the various species, concluded that *C. undulatus, C. pachyderma,* and *C. parryi* are distinct species. (Similar habitats have resulted in the convergent similarity between *C. parryi* and *C. undulatus*, confusing to many taxonomists.) They further believed that *C. undulatus* became extinct over the territory from the right bank of the Lena River to Alaska and Canada during Sangamon time. In early Wisconsinan time, *C. richardsoni* migrated from the south and by late Wisconsinan time *C. parryi* had originated from it (Liapunova and Vorontsov, 1970, p. 1037).

A detailed study (Repenning and others, 1964) of fossil rodents from the valley-bottom silt deposits of Wisconsinan and post-Wisconsinan age near Tofty, 150 km west of Fairbanks, listed the following:

Lepus sp
Spermophilus (*Spermophilus*) *undulatus* (Pallas)
Beaver (gnawed wood)
Discrostonyx torquatus
Lemmus sibericus
Synaptomys sp. (identified by T. M. Stout)
Clethrionomys sp. (identified by T. M. Stout)
Microtus miurus
Microtus (large form)

Also present in the Wisconsinan silt near Tofty are remains of Proboscidea?, *Equus, Rangifer*(?), *Alces*(?), *Bison,* bovid, and an unidentified fish.

Most of the mammal remains from the Fairbanks area are from retransported valley-bottom silt of Wisconsinan age rich in organic material (figs. 20, 29; table

TABLE 12.—*Fossil mammals reported from perennially frozen creek-valley silt deposits of Pleistocene age in the Fairbanks district*

Name	Genus
Insectivora:	
Shrew	Genus not determined.[1]
Carnivora:	
Bear	*Ursus* sp.[2]
Do	*Ursus yukonensis* Lambe.[3]
Do	*Ursus arctos.*[4]
Dire wolf	*Canis dirus alaskensis* Frick.[3]
Coyote	*Canis latrans.*[4]
Wolf	*Canis* sp.[5]
Do	*Canis lupus.*[4]
Fox	*Vulpes* sp.[2]
Badger	*Taxidea* sp.[2]
Weasellike animal	Genus not determined.[2]
Wolverine	*Gulo* sp.[5]
Saber-toothed tiger	*Smilodon* sp.[2,4]
Lion	*Felis atrox alaskensis* Frick.[3]
Lynx	*Felis* (*lynx*) sp.[5]
Proboscidea:	
Elephant	*Mammuthus primigenius.*[6,7]
Mastodon	*Mastodon* sp.[2]
Do	*Mastodon americanus alaskensis.*[6]
Perissodactyla:	
Horse	*Equus alaskae Hay.*[3]
Do	*Equus lambei Hay.*[8]
Do	*Equus caballus.*[4]
Artiodactyla:	
Camel	*Camelops race* Gidley.[3]
Antelope	*Saiga* sp.[2,9]
Do	*Saiga ricei* Frick.[10,9]
Do	*Saiga tatarica.*[4]
Bison	*Bison* (*Superbison*) *crassicornis* Richardson.[11]
Do	*Bison* (*Bison*) *preoccidentalis* S. and K.[11]
Do	*Bison* (*Platycerobison*) *alaskensis* Rhoads.[11]
Artiodactyla—Continued	
Bison	*Bison* (*Platycerobison*) *geisti* S. and K.[11]
Caribou	*Rangifer* sp.[5]
Do	*Rangifer tarandus.*[4]
Moose	*Alces* sp.[5]
Do	*Alces alces.*[4]
Elk	*Cervus* sp.[2]
Do	*Cervus elaphus.*[4]
Elk, giant	*Cervalces alaskensis* Frick.[10]
Sheep	*Ovis dalli kaiseni* Frick.[10]
Do	*Ovis dorshi Frick.*[10]
Do	*Ovis nivicola*[4]
Musk ox	*Ovibos yukonensis* Gidley.[10]
Do	(?)*Ovibos giganteus* Frick.[10]
Do	*Ovibos moschatus.*[4]
Do	*Praeovibos.*[12]
Musk oxlike Bovide	*Symbos tyrrelli* Osgood.[10]
Do	*Bootherium nivicolens* Hay.[10]
Yaklike Bovid	*Bos* sp.[4]
Do	*Bos* (*Poephagus*) *bunnelli* Frick.[10]
Edentata:	
Groundsloth	*Megalonyx* sp.[13]
Rodentia:	
Beaver	*Castor* sp.[5]
Ground squirrel	*Citellus parryi.*[14,15,16]
Vole	*Microtus* sp.[5]
Do	*Microtus gregalis.*[17]
Lemming	*Lemmus* sp.[2]
Do	*Lemmus sibericus.*[17]
Collared lemming	*Dicrostonyx.*[2]
Do	*Dicrostonyx torquatus.*[17]
Porcupine	*Erethizon* sp.[5]
Lagomorpha:	
Hare	*Lepus* sp.[5]
Pika	*Ochotona whartoni.*[18]

[1]Pruitt (1955) (oral commun.; Arctic Aereomedical Lab., U.S. Air Force, Fairbanks, Alaska).
[2]Geist (1947–55) (oral commun., research associate, Dept. of Paleontology, Univ. Alaska).
[3]Frick (1930).
[4]Guthrie (1968a).
[5]Geist (1953).
[6]Frick (1933).
[7]Osborn (1942).
[8]Mertie (1937).
[9]Skinner (1970) (written commun., March 3, 1970) Am. Mus. Nat. History.
[10]Frick (1937).
[11]Skinner and Kaisen (1947); referred to as *Bison Priscus* by Guthrie (1966a, p. 739).
[12]Ray (1967), (written commun., Oct. 25, 1967) Supervisor, Div. Vertebrate Paleontology, Smithsonian Institution.
[13]Stock (1942).
[14]Hill (1942).
[15]Stout (1956) (written commun., Aug. 12, 1956) Univ. Nebraska.
[16]Guthrie (1968b) used the term "*Citellus undulatus*". (See footnote 5 and Rausch, 1953, p. 121.)
[17]Guthrie (1968b).
[18]Guthrie and Matthews (1971).

12). Most of the fossils are found in valley bottoms, owing to gradual downslope movement. Some transportation also occurs down the stream axes, and the greatest concentrations are found where small tributaries join large creeks. These great accumulations of bones are thus not "animal cemeteries" or unnatural concentrations. Bones are recovered throughout a given valley and do not come only from "bone pits" as stressed by Osborn (1942, p. 1035). Most vertebrate remains in interior Alaska are Wisconsinan; the vertebrate-rich valley-bottom facies of the loess of Illinoian age was, in most instances, removed before Wisconsinan sediments were deposited (fig. 20).

The writer believes that all taxa reported from the perennially frozen silt in the Fairbanks area (table 11) are represented in silt of Wisconsinan age (fig. 42), except perhaps *Praeovibos*, because in most instances only sediments of Wisconsinan age were exposed in mining cuts at the times of collection.

Carcasses

Frozen mammal carcasses have been known from the Arctic and subarctic for centuries (Nordenskiold, 1881, p. 408–410; Herz, 1904; Tolmachoff, 1929; Markov and Popov, 1959; Tikhomirov, 1958). The first direct report of a carcass of a mammoth in frozen ground of Siberia was by E. Yssbrants Ides in 1692 (Tolmachoff, 1929, p. 21).

Although not so well known as the Siberian carcasses, partial remains consisting of bones with dried flesh, skin, or hair still clinging to them have been found in frozen silt (Dall, 1896, p. 857; Quackenbush, 1909, p. 108–109; Wilderson, 1932, p. 6; Skarland, 1949, p. 68; Geist, 1940, 1953; Péwé, 1966c). Many important finds of carcasses have been from the retransported organic-rich silt of Wisconsinan age in the Fairbanks area. Partial carcasses of mammoth, bison, musk ox, *Symbos*, *Bootherium*, moose, horse, lynx, caribou, and ground squirrel were found in the environs of Fairbanks (Péwé, 1952a, p. 123–126). Several complete carcasses of the ground squirrel have been found in nests. The most celebrated find was the partial forequarters of a baby mammoth, collected in 1948 from Fairbanks Creek (fig. 43) (Anthony, 1949). The well-preserved hide of the head, neck, trunk, and one front leg was about 6 mm thick and almost hairless. The partial carcass was preserved in commercial glycerine by the collector, O. W.

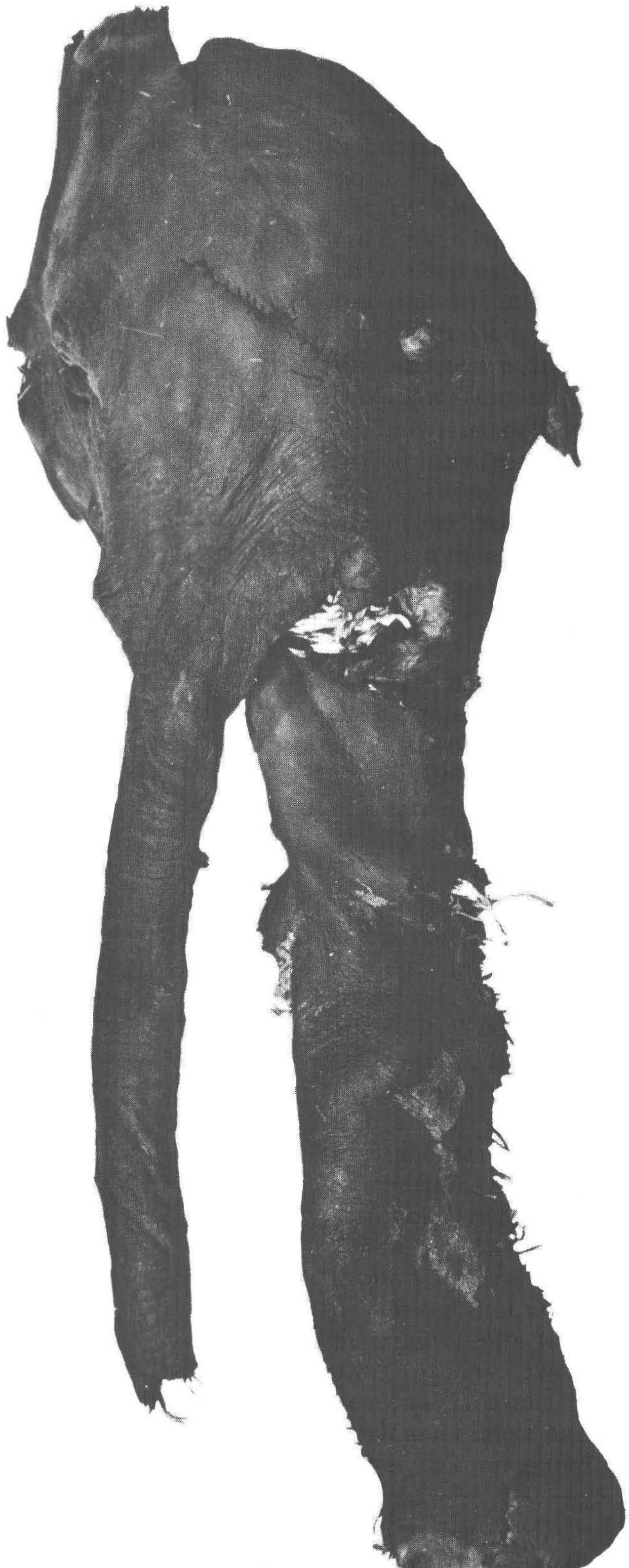


FIGURE 43.—Partial carcass of a juvenile mammoth recovered in 1948 by Otto W. Geist from perennially frozen silt of Wisconsinan age at Fairbanks Creek, 30 km northeast of Fairbanks, Alaska. Photograph from Bunnell Collection, University of Alaska.

TABLE 13.—*Radiocarbon dates of Pleistocene mammals from the perennially frozen silt in the Fairbanks area*

[Specimens collected by O. W. Geist were provisionally identified by Geist; they were then placed in the Frick Collection of the American Museum of Natural History and identification was confirmed by members of the staff of the American Museum. Laboratory numbers: ST = radiocarbon laboratory of Geological Survey of Sweden, Stockholm; SI = radiocarbon laboratory of Smithsonian Institution, Washington, D.C.; GX = radiocarbon laboratory of Geochron, Inc., Boston; M = radiocarbon laboratory of University of Michigan, Ann Arbor; L = radiocarbon laboratory of Lamont Geological Observatory, New York; W = radiocarbon laboratory of U.S. Geological Survey, Washington, D.C.]

Animal	Sample	Age years (B.P.)	Laboratory No.	Collector and year	Locality	Submitter	Source	Remarks
Carnivora:								
Felis atrox	Tendon from left tibia.	22,680±300	SI 456	O. W. Geist, 1938.	Ester Creek.	C. E. Ray	Stuckenrath and Mielke (1970, p. 203).	
Proboscidea:								
Mammuthus primigenius.	Flesh from lower leg.	15,380±300	SI 453	R. H. Osborne, 1940.	Fairbanks Creek.	R. H. Osborne	do	85 ft below surface; partial carcass.
sp	Hide from carcass.	21,300±1300	L 601	O. W. Geist, 1948.	do	W. R. Farrand	Farrand (1961).	Hide soaked in glycerine by collector; date invalid(?).
Do	Hair from skull.	32,700±980	ST 1632	O. W. Geist, 1951.	Dome Creek	T. L. Péwé	Péwé (1965a, p. 33).	
Perissodactyla:								
Equus	Bone	26,760±300	SI 355	H. L. Foster, 1965.	Lost Chicken Creek.	F. C. Whitmore, Jr.	Mielke and Long (1969, p. 177).	
Artiodactyla:								
Bison (Bison) preoccidentalis.	Horn sheath	11,735±130	ST 1631	O. W. Geist, 1937.	Cleary Creek	T. L. Péwé	Péwé (1965a, p. 33).	From internal piece of type specimen;
Do	do	12,460±320	SI 290	do	do	C. E. Ray	Mielke and Long (1969, p. 177).	Skinner and Kaisen (1947, p. 174–176).
Bison	do	>35,000	SI 844	O. W. Geist, 1938.	Little Eldorado Creek.	R. D. Guthrie	Stuckenrath and Mielke (1973, p. 397).	
preoccidentalis.	do	5340±110	SI 845	O. W. Geist, 1939.	Goldstream area.	do	Stuckenrath and Mielke (1973, p. 397).	Not *preoccidentalis* or contamination present
Bison	do	29,295±2440	SI 842	O. W. Geist, 1940.	Cripple Creek.	do	Stuckenrath and Mielke (1973, p. 397).	Small sample, diluted
Do	do	>39,000	SI 840	O. W. Geist, 1947.	do	do	Stuckenrath and Mielke (1973, p. 396).	
Do	do	21,065±1365	SI 839	do	do	do	Stuckenrath and Mielke (1973, p. 396).	
Do	do	18,000±200	SI 841	O. W. Geist, 1948.	Manley Hot Springs.	do	Stuckenrath and Mielke (1973, p. 397).	
(*Superbison*) *crassicornis.*	Hide from carcass.	>28,000	L 127	O. W. Geist, 1951.	Dome Creek	O. W. Geist	Kulp and others, (1952, p. 411).	
Do	Hide and hair from carcass.	31,400±2040	ST 1721	do	do	T. L. Péwé	Péwé (1965a, p. 33).	
Bison	Horn sheath	17,170±840	SI 838	O. W. Geist, 1952.	Fairbanks Creek.	R. D. Guthrie	Stuckenrath and Mielke (1973, p. 396).	
Do	do	20,445±885	SI 837	do	do	do	do	
preoccidentalis.	do	31,980±4490	SI 843	O. W. Geist, no date.	Fairbanks	do	do	Do.
(*Superbison*) *crassicornis.*	do	16,400±2000	M 38	do	Creek near Fairbanks.	C. Hibbard	Crane (1956 p. 670).	Date obtained by solid carbon method.
(female).	Hide from carcass.	11,980±135	ST 1633	do	Fairbanks Creek.	T. L. Péwé	Péwé (1965a, p. 33).	
Ovibos sp	Hair from hind limb	17,210±500	SI 454	O. W. Geist, 1940.	Fairbanks Creek.	C. E. Ray	Stuckenrath and Mielke (1970, p. 203).	
	Muscle from scalp.	24,140±2200	SI 455	do	do	do	do	From small sample, diluted.
Symbos	Horn sheath.	25,090±1070	SI 850	O. W. Geist, 1939.	Cleary Creek.	R. D. Guthrie	Stuckenrath and Mielke (1973, p. 397).	Do.
giganteus	Dung(?)	>40,000	SI 291	do	Creek near Fairbanks.	C. E. Ray	Mielke and Long (1969 p. 177).	Date may be on gut.
Symbos	Horn sheath.	17,695±445	SI 851	O. W. Geist, 1952.	Dome Creek.	R. D. Guthrie	Stuckenrath and Mielke, (1973, p. 397).	
Bootherium nivicoleus.	do	22,540±900	SI 292	O. W. Geist, 1935.	Creek near Fairbanks.	C. E. Ray and M. F. Skinner.	Long and Mielke (1967, p. 380).	
Rodentia:								
Citellus undulatus.	Nest	14,760±850	GX 0250	T. L. Péwé 1963.	Chatanika River at Livengood Road.	T. L. Péwé	Péwé (1965a, p. 35).	
Do	Coprolites	14,510±450	W 2703	T. L. Péwé, 1971	do	do	M. Rubin (oral commun., 1972).	

Geist; therefore the date of 21,300 ± 1,300 years must, unfortunately, be questioned (table 13).

A partial bison carcass was found in perennially frozen silt at Dome Creek in 1951 during mining operations. The carcass consists of a head, complete with hide, horns, and one ear, as well as four legs with hooves;

FIGURE 44.—Partial carcass of an extinct bison, *Bison (Superbison) crassicornis*, discovered August 1951, in perennially frozen, retransported, organic-rich silt of Wisconsinan age during placer gold mining operations on Dome Creek, 13 miles north of Fairbanks, Alaska. Radiocarbon dated at 31,400 (+2,040 or −1,815) years old (ST 1721). Collector, O. W. Geist, is examining fragment of the hide. Photograph No. 600 by T. L. Péwé, September 3, 1951.

much of the torso hide is about 4 mm thick (fig. 44). The carcass shows evidence of having been transported only a short distance. A date of more than 28,000 years old (L–217) was obtained on a piece of the carcass in 1951 (table 13). In 1965 a date of 31,400±2,040 years old (ST 1721) was obtained by the radiocarbon laboratory at the Geological Survey of Sweden on a piece of the same carcass (Péwé, 1965a). Pieces of fur and hide of a female superbison recovered from Fairbanks Creek near Fairbanks have been dated as 11,950±135 years old (ST 1633) (Péwé, 1965a).

At Dome Creek near Fairbanks, a fairly complete skull and tusks of mammoth were recovered in the early 1950's. The mammoth tusks are 3.7 m long (outside curve) and weigh approximately 160 kg apiece. Preserved with the skull was considerable mammoth hair which was dated at 32,700±980 years old (ST 1632) (Péwé, 1965c). The writer submitted some of this hair to J. M. Gillespie, Senior Principal Research Scientist of Division of Protein Chemistry of CISRO, Australia, for examination. He reported (Gillespie, 1970) that although the mammoth hair at first sight seems undamaged, is intact externally, and is quite strong, it has, in fact, been extensively damaged, perhaps during decomposition that preceded freezing. This is additional evidence against instantaneous death and refrigeration.

Some of the hair was also submitted to Prof. Shoichi Yada of the Department of Legal Medicine at Gifu University, School of Medicine, Japan. He reported (written commun., May 13, 1971) that, morphologically, most of the specimens were in good condition, presumably owing to the extremely high resistance of the keratin tissue to decomposition. He further stated that the mammoth hair is furnished with an A antigen similar to that of group A human hair and mentioned it is surprising indeed that the blood group A antigen occurring in the mammoth hair had successfully survived through tens of milleniums of weathering without losing its serological specificity.

Within the last few years, samples of fleshy organic matter clinging to bones collected in the Fairbanks area in the 1930's and 1940's have been dated, especially by the Smithsonian Institution (table 13). R. D. Guthrie is continuing this work and is currently having approximately 100 mammal specimens from the Fairbanks area radiocarbon dated by the Smithsonian Institution.

Many partial carcasses of rhinoceroses were reported from Quaternary deposits in Siberia (Tolmachoff, 1929, p. 20), but no carcasses or even bones of this animal have been recognized in Alaska. Flerow (1967, p. 273) believed the rhinoceroses did not migrate to Alaska because they were forest inhabitants and probably could not feed on the grass of the Bering land bridge.

The frozen carcasses of extinct animals have long caught the fancy of the general public and scientist alike, especially the hugh mammoths reported from remote areas of Siberia. Stories quickly sprang up and have been repeated for many years, both in print and orally, regarding the mammoth meat being served at banquets, mammoths dying with buttercups in their mouths, and the large beasts perishing as cataclysmic climatic changes took place turning the tropical climate into the frigid Arctic. It is not necessary to envision such conditions to explain the distribution, preservation, and extinction of the ice age vertebrates as suggested over the years by Howorth (1887), Velikovsky (1955), Hapgood (1958, 1960), Hooker (1958), Sanderson (1960), Patten (1966), and Dring (1967).

The presence of vertebrates of many species in the sediments of Illinoian and Wisconsinan age in Alaska abolishes the long-held idea that the animals must have lived only in interglacial times (Flint, 1957, 1971) or were exterminated by severe conditions of glacial times (Johnston, 1933, p. 32). Detailed work, especially near Fairbanks, demonstrated that all the carcasses are between 10,000 and 70,000 years old and could hardly be older because they probably would not have survived the Sangamon Interglaciation, a time when permafrost probably disappeared in central Alaska (Péwé, 1952a, 1958b). It is evident that most of the permafrost present today in the Fairbanks area must have existed since Wisconsinan times to preserve the carcasses in nature's deep freeze. The geologic association, plus radiocarbon dates (table 13), indicates that the carcasses are Wisconsinan in age and not "a million years old," as casually announced.

The mammoth, bison, and others lived on the tundra-covered hills and grassy flood plains (fig. 42), when a rigorous glacial climate prevailed. The tundra then, as today, supported several species of *Ranunculus* (buttercup) and *Potentilla*. For example, a fossil seed cache of *Citellus undulatus* collected by the writer from Wisconsinan sediments in Dome Creek near Fairbanks was found by R. W. Browne (written commun., Jan. 29, 1953) to consist of seeds of *Ranunculus hyperboreus* Rottboell (Arctic buttercup) and *Potentilla* sp. (a cinquefoil). Buttercups also made up a part of the diet of mammoths; pollen of *Ranunculus* has been found between the teeth and in the stomach of the frozen carcass of the Berezovka mammoth. Some mammoths did indeed die "with buttercups in their mouths," but this is a perfectly natural association and occurrence. Virtually all evidence suggests that the carcasses represent natural deaths in a rigorous environment (Farrand, 1961).

HOLOCENE

In the Fairbanks area no bones of extinct vertebrates have been found in the post-Wisconsinan sediments (figs. 20, 29). Only bones of living forms have been found. A complete frozen and untransported moose carcass was found in silt of Holocene age (Giddings, oral commun., Nov. 4, 1950). Guthrie (1968b) reported *Microtus xanthognathus,* a living form, from the frozen Holocene sediments at Eva Creek. Hopkins (1963) stated that there is no evidence for extinct species living after 11,000 years ago in the Seward Peninsula. He recorded the presence of beaver (*Castor* sp.) in central Seward Peninsula between 8,000 and 9,500 years ago and again around 3,500 years ago; they do not live there today. McCulloch reported an expanded range of beaver at about the same time in the Kotzebue Sound area (McCulloch and Hopkins, 1966).

In the Holocene flood plain deposits of large streams, rare finds of *Bison bison* have been made. Geist stated (oral commun., 1950) that the finds of *Bison bison* reported in Skinner and Kaisen (1947) were from such sediments and that a bone of *Bison bison* was found in Holocene sediments of the Chena River (pl.1) at Fairbanks.

The return of mammals to areas glaciated in Wisconsinan time, generally referred to as the postglacial distribution pattern of mammals, has been studied in temperate latitudes. However, except for Klein's work (1965) in the southern coastal regions, Karlstrom's report (1969) on Kodiak Island, and the studies of the distribution of early man in Holocene time, little systematic work has been done on this problem in Alaska. Perhaps the migration of mammals locally north and south from the large interior refugium presents no problem. Nevertheless, the present land areas of the coastal regions of Alaska bordering the Gulf of Alaska were virtually completely glaciated, and the present flora and fauna of the region had to become established in the last 10,000 years, since the recession of the ice. Klein (1965) carefully showed that although some local refugia existed in this region, all large mammals in the region either migrated through mountain passes from the refugium in the interior of Alaska or came from south of the ice sheet. For example, the brown and black bear, wolf, goat, and mountain sheep probably came from the north as did one form of moose. Other moose forms and the blacktailed deer are known to have migrated from the south.

EXTINCTION

In Alaska, as in the rest of the world, many mammals became extinct at the end of Wisconsinan time. Moreover, many large mammal genera, grazers, browsers, and predators, present in both Illinoian and Wisconsinan time, became extinct. In Alaska the most common and largest were the bison, mammoth, and horse. These three grazers represent more than 90 percent of the large-mammal biomass (Guthrie, 1968a). Guthrie (1968a) concluded that during late Pleistocene time, vegetational patterns must have been radically different from those today, especially in central Alaska. He believed that a widespread grassland environment existed in the northern Holarctic region, rather than a shrub tundra or herbaceous tundra environment as exists today, and that it was the disappearance of this grassland that caused the animals to become extinct.

It has long been known from studies of pollen and presence of grazers that tree line was lowered 450–600 m in central Alaska and that the tundra environment was much more widespread than now. However, pollen studies by Livingstone (1955), Colinvaux (1967a), and Heusser (1963b) in western and northern Alaska did not indicate a "grassland" but, rather, an herbaceous to shrub tundra with some grass. Pollen analyses by E. C. Barghoorn on samples collected by the writer from critical localities in Illinoian and Wisconsinan sediments near Fairbanks indicated an absence of coniferous trees and the presence of a grass-sedge-moss cover with dwarf birch and willow, but no "grassland."

A simple model of vegetation distribution in central Alaska during late Pleistocene time may account for extinction of a large percentage of grazers, in part without the necessity of postulating unique climatic conditions. During Illinoian and Wisconsinan time, all major and most minor streams in Alaska drained glaciated areas. These streams were extensively braided with wide, relatively flat flood plains, and on the flood plains vegetation was in a disclimax condition. Such widespread, flat flood plains also existed on the Bering land bridge. Evidence for the existence of these nonforested, wide, braided flood plains is the abundant loess of Illinoian and Wisconsinan age in Alaska adjacent to the flood plains (figs. 16, 20) (Péwé, 1951a, 1968a; Péwé and Holmes, 1964; Péwé and others 1966); the flood plains were the source of this loess (fig. 42).

These extensive flood plains and adjacent low terraces in central, western, and northern Alaska were ideal locations for grasslands in Pleistocene time. The soil was well drained, permafrost table low, and muskeg at a minimum. Today the greatly restricted flood plains of glacial streams support grasslands. The two places where modern bison have successfully been introduced in central Alaska are in such habitats, and, significantly, extinction of the few lingering bison (*Bison bison*) of the great Pleistocene herds is recorded only in Holocene sediments of large glacial flood plains, and not in the creek valleys of the uplands where the Pleistocene bison remains are found.

The wide grassy flood plains of major rivers were the home of and best food supply for the late Pleistocene grazers, especially in winter. In summer, undoubtedly many, if not most, grazers wandered to the hills, to the less grassy tundra, perhaps to escape flies. Like the modern elk (Murie, 1951, p. 241), perhaps the Pleistocene grazers ate less grass and more sedge, herbs, and other plants in summer, but in winter the grass, perhaps always their choice, became a necessity. Horses used by miners, trappers, and geologists years ago in Alaska subsisted on "native grasses" during the summer. But when the first frost came in the middle of September, the horse feed was "ruined"; the plants, mainly nongrass subsistence, lost their nutritional value (Brooks, 1953, p. 408). If this sort of seasonal change in environment also took place in early Holocene time, the grazers could live after a fashion away from the flood plain, especially in summer, but in the winter they probably searched for grass on the windswept flood plains where the snow blanket was thinnest.

Therefore, the large population of grazers with fewer browsers and predators evidently thrived during the rigorous glacial climate in the nonforested areas of Alaska. In interglacial times forests reinvaded the valleys and low hills, and an environment not too different from that today must have existed in central Alaska. This tremendous reduction of their ideal habitat must have caused extensive stress for the grazers. Still, they did not become extinct in the Yarmouth(?) or Sangamon Interglaciations but died out at the end of the Wisconsinan Stage.

If the grazers did not become extinct during the time of stress during interglacial times, it is unlikely that they would become extinct at the end of the Wisconsinan because of the loss of grassy habitat alone. No doubt an additional stress was added at this time, in the form of man. The combination of loss of ideal habitat and more efficient predation by man is therefore regarded as having caused the demise of the late Pleistocene grazers in Alaska, 10,000 years ago.

MARINE VETEBRATE AND BIRD REMAINS

To complete a summary of vertebrate remains from Quaternary deposits of Alaska, mention should be made of several finds that may lead to a detailed study. A walrus tusk and fish remains were recovered from the estuarine silt of Sangamon age (Pelukian transgression) at Nome (Hopkins and others, 1960, p. 52), and a metacarpal of a walrus (*Odobenus*) was found in late Pleistocene sediments from the south shore of Kuk Inlet on the north coast of Alaska (William Quaide, unpub. data). McCulloch dug a walrus humerus out of an old beach ridge 5–6 miles up the Kokolik River, northeast of Point Lay, and at one time he considered these to be deposits of the Pelukian transgression. Also, spinal disks from whales were collected by O. W. Geist (oral commun., 1959) from the "muck" banks of the Inglutalik River (pl. 1), where it empties into Norton Bay at the base of Seward Peninsula. The first discovery of Steller's sea cow (*Hydrodamalis*) in place in Pleistocene deposits was made in beach sand and gravel thought to be interglacial, 35 m above present sea level on Amchitka, Aleutian Islands (Gard and others, 1972). The bone was dated at about 135,000 years old and is thought to be Sangamon in age.

C. A. Repenning (written commun., Apr. 20, 1970) supplied the following identification of marine vertebrates of Pleistocene age collection by D. M. Hopkins in western Alaska from various Pleistocene horizons (table 10):

Phoca vitalina(?)
(?)Callorhinus sp.
Pusa hispida
Eumetopias sp.

In additon, F. H. Fay (written commun., Apr. 24, 1970) stated that bones of walrus of Pleistocene age were recovered from Nelson and St. Lawrence Islands.

Rare bones of birds of Pleistocene age were reported from the Fairbanks are (O. W. Geist, written commun., Dec. 5, 1950) and from the Arctic Coastal Plain. No informaion is available as to the exact age and identification. A unique study by Hoskin, Guthrie, and Hoffman (1970) demonstrated that bird gastroliths of Pleistocene age are perhaps widespread in Alaska. It would be well for workers in the Quaternary deposits of Alaska to be aware of the existence and the properties of these small gastroliths.

SUMMARY

Remains of Pleistocene mammals are widespread in Alaska, and large collections have been made. Although stratigraphic context is unknown for most of the early collections, good stratigraphic control is available for later and current work. Most of the remains are of Wisconsinan age, but there is a growing number of significant pre-Wisconsinan specimens. Only one good early Pleistocene collection is known. For a better understanding of mammal life in Alaska, more detailed knowledge is needed concerning early Pleistocene time and especially regarding the time between early Pleistocene and Illinoian time. Detailed meticulous sieving of the extensive silt deposits, especially in valley bottoms, should continue to yield valuable results.

It is abundantly clear from the work in Alaska over the last 20 years that many taxa were present in North America earlier than generally recognized. Alaska perhaps will provide a significant link in the understanding of the time of Pleistocene vertebrate origins in North America.

Remnants of frozen carcasses of Pleistocene mammals in permafrost are not rare in unglaciated Alaska, especially central Alaska. Geologic association plus radiocarbon dates indicate all carcasses are Wisconsinan in age and thus abolish the long-held idea that the animals must have lived only in interglacial times. All specimens show some sign of decay, and the animals were not frozen instantaneously by cataclysmic happenings. Specimens are available for detailed laboratory studies, and it is hoped that more work on the preserved specimens will be initiated by scientists in various disciplines.

Studies of Alaska vertebrates shed light on extinction of mammals at the end of the Pleistocene. Although there is no complete agreement in regard to details, it is thought that the loss of grassy habitat and increased predation by man may have caused the extinction of the late Pleistocene grazers in Alaska 10,000 years ago.

CLIMATE

Quaternary climatic fluctuations in Alaska as elsewhere in the world were reponsible for formation and disappearance of glaciers and permafrost and changes in distribution of plants and animals. In this review the climatic changes of late Cenozoic time, mainly Quaternary, are based on available physical and biological data from glaciated and unglaciated areas. The causes of the climatic changes of the Quaternary are not dealt with here; they were recently summarized by Mitchell (1965; 1968).

EVIDENCE FROM GLACIATED AREAS

A parameter to be considered in evaluating past Quaternary climates in the glaciated areas, in addition to the existence and size of the glacial advances, is the past position of snowline and what it may mean in light of the vertical change of temperature of the atmosphere (lapse rate). A probable Pleistocene depression of mean annual or mean monthly air temperatures can be calculated by determining the difference in elevation between the present (fig. 11) and Pleistocene (fig. 12) snowlines and by multiplying this figure by the current average adiabatic lapse rate. Basic assumptions that must be made are that the Pleistocene lapse rate were similar to the present and that temperature decrease alone is responsible for lowering the snowline. Calculation of temperature depression from past snowline positions in Quaternary time has only been undertaken in Alaska in the last decade (Péwé and Burbank, 1960; Porter, 1966; Péwé and others, 1967; Péwé and others, 1969). This technique is not applicable to climatic changes earlier than Illinoian time because cirques of pre-Illinoian age are not recognized.

The temperature decrease calculated by this method is necessarily a minimum figure because snowline on even small valley glaciers is lower than cirque base. Probably more accurate is a calculation based on determining the position of the equilibrium line of past healthy glaciers using the accumulation area ratio (Porter, 1968, 1970). This technique has not yet been applied to Alaska except at Indian Mountain (R. D. Reger and T. L. Péwé, unpub. data, 1974).

EVIDENCE FROM UNGLACIATED AREAS

Geologic features useful in detecting Quaternary climatic changes in a periglacial area are frost features such as permafrost, ice wedges, altiplanation terraces, solifluction deposits, and patterned ground, and other features. The use of such features in interpretation of Quaternary climates in temperate latitudes is widespread (Wright, 1961); however, only recently have quantitative data become available to permit a more thorough understanding of the periglacial environment.

The mere presence or absence of permafrost as a climatic indicator is actually less valuable in Alaska than in temperate latitudes because permafrost is still present in most of Alaska and is actively forming in certain areas where the mean annual air temperature is −1°C or colder. The presence or absence of ice wedges is more valuable in Alaska than in temperate latitudes because of the different types of ice wedges still present, such as active, inactive, and ice wedge pseudomorphs.

Ice wedges, or foliated ice masses, constitute the critical parameter in permafrost that is of the most value in reconstruction of past climates. From the earlier discussion of permafrost, it is apparent that ice wedges readily grow in regions where the mean annual air temperature is colder than −6°C to −8°C. Ice wedges may exist inactively for thousands of years in areas where the mean annual air temperature is between −2° and −6°C. Of course, it is always possibile that cold microclimatic conditions may occur so that a few small local ice wedges could grow in a broad area of "inactive" ice wedges.

Other geologic features used as climatic indicators are loess, sand dunes, and ventifacts. In Alaska these features (figs. 15, 16, 17) do not necessarily indicate greater aridity or wind action than at present, nor are they in themselves indications of a colder climate. It is felt that widespread loess and sand dune deposits of Quaternary age are mainly indicators of much larger source areas of silt and sand in the past. All these features, at least in central Alaska, are associated with extensive glaciations and, therefore, form with the changes of the climate.

Biogeographic evidences of past climates are the changes in the position of tree line and floral communities (from pollen analyses) and the changes in environment and distribution of marine and terrestrial fauna.

LATE TERTIARY

As a generalization, it is believed that the climate of late Tertiary time in Alaska was relatively equable and not intensely cold, although glaciers were present in Alaska 10–13 m.y. ago (table 1).

Glaciation was probably extensive in the mountains facing the Gulf of Alaska. Ice pushed to the interior from the Wrangell and St. Elias Mountains (Denton and Armstrong, 1969). Ice reached the sea to the south and sporadically for at least 10 m.y. repeatedly formed shelf ice and discharged icebergs into the ocean (Taliaferro, 1932; Miller, 1953a, 1957; Plafker, 1971; Bandy and others, 1969; Kent and others, 1971; Scholl and others, 1971). A record of the fauna and the glacial deposits exists in the Yakataga Formation and shows that the adjacent sea was about 10°C colder than now as indicated by the presence of the cold-water planktonic foraminifer *Turborotalia pachyderma* (Bandy and others, 1969). Plafker (written commun., Apr. 21, 1970) believed the molluscan and foraminiferal faunas indicate a decrease of surface-water temperature of 10°–15°C in early Yakataga time.

The Yakataga Formation is in an extremely rugged (pl. 1) and tectonically active region which today receives exceptionally heavy precipitation (upwards of 250 cm per year at sea level) (fig. 2). Despite a relatively mild climate, it is one of the most heavily glaciated areas in Alaska, and many glaciers reach sea level, invading the temperate Sitka spruce and hemlock forest that fringes the shore. The presence in this area of late Tertiary deposits of glacial origin probably indicates that relief was high, rainfall heavy, and the climate cool-temperate, much like the present, and not intensely cold.

Leaves of beech (*Fagus* sp.) were found with an early Pliocene mollusk fauna in the fine-grained beds interstratified with conglomeratic sandy mudstone at Cenotaph Island in Lituya Bay, indicating that land climates at sea level were somewhat milder than at present and that the glaciers probably invaded a forest of beech trees (Hopkins, 1972).

From the Arctic Basin comes information regarding late Tertiary climate based on interpretation of fauna and sediments in deep cores. Evidence for glaciation and cooling, at least in the adjacent highlands 4–6 m.y. ago, is indicated by ice-rafted pebbles. This does not indicate ice-pack conditions or especially cold waters. According to Hopkins (1972), the Arctic Ocean could not have been frozen or even been very cold about 3.5 m.y. ago, because at that time Pacific mollusks invaded the North Atlantic (Beringian transgression) (table 1). He further stated that the recent discovery of *Turritella* in Pliocene marine beds offshore at Nome indicates temperate shelf waters during Pliocene time. Later, ice-rafting increased and pack ice cover occurred (table 1).

The forest vegetation of spruce, birch, pine, and hemlock in northern Seward Peninsula 5.7 m.y. ago (Hopkins and others, 1971) (table 1) indicates an environment similar to southeastern Alaska and coastal British Columbia rather than the Arctic climate of today. In general, flora of Beringian times (table 1) indicates that the climate of western and northern Alaska was considerably warmer than at present (Hopkins, 1967a, p. 61). The tundra cover and associated cooler climate probably did not arise in the northern Seward Peninsula until early Pleistocene time.

EARLY PLEISTOCENE

The early and middle Pleistocene of Alaska record glacial episodes, interglaciations, and major changes in climate with corresponding modification of permafrost and other periglacial features and changes in the distribution of the flora and fauna of the sea and land. In this report, climatic changes of pre-Yarmouth time are labeled as early and middle Pleistocene in age, and provincial names are used.

A cold episode of early and middle Pleistocene age is recorded in many areas of Alaska. The cooling of the climate resulted in the Iron Creek Glaciation on Seward Peninsula, deposits of which underlie beds of the Anvilian transgression and are thought by Hopkins to be 1.0±0.5 m.y. old. The transgression is less than 1.8 m.y. old but more than 700,000 years old. Glacial advances elsewhere in Alaska around this time or before may be the Browne Glaciation (Wahrhaftig, 1958) in the Alaska Range, thought to be more than 2.7 m.y. old (Wahrhaftig, oral commun., July 4, 1970), and the Mount Susitna Glaciation in Cook Inlet (Karlstrom, 1964).

An excellent record of tundra biota is reported from Cape Deceit on the south shore of Kotzebue Sound (Guthrie and Matthews, 1971). Ice wedge pseudomorphs are reported at this locality, and Hopkins (1972) believed they represent a permafrost period of 1.0 ±0.5 m.y. ago.

In valleys of small streams and rivers in central Alaska, an angular coarse auriferous gravel of great antiquity is preserved on benches (fig. 20) and represents a drainage pattern slightly different from that of the present. In places, the material grades laterally up slopes into solifluction deposits. The gravel is the result of the accumulation of frost-rived local debris produced in a rigorous climate. The solifluction layer is widespread in central Alaska and contains well-developed ice wedge pseudomorphs (fig. 33). They are thought to be at least 1 m.y. old inasmuch as they represent the older of two permafrost periods, both of which predate the Illinoian glaciation. As outlined earlier, ice wedges indicate that the mean annual air temperature was colder than −6° to −8°C.

Data regarding temperature of sea water in early Pleistocene time are rare, but mollusks (particularly *Portlandia*) dredged from rocks of early Pleistocene age in the southern Bering Sea are indicative of water temperatures that were colder than at present (Hopkins and others, 1969). (*Portlandia* is not known to occur south of the Bering Strait at the present time.)

After the early Pleistocene cold period(s), a warmer period termed "the Anvilian transgression" occurred. Hopkins (1967b) stated that mollusks indicate a warmer sea water temperature than in the modern Bering and Chukchi Seas as well as the Arctic Ocean (Hopkins, 1972). For example, *Natica janthostoma* was found in Anvilian sediments near Nome and Skull Cliff in northern Alaska but now is limited to waters adjoining Japan. During the warmer part of the Anvilian transgression, winter ice probably did not reach Nome and may not have reached the Arctic coast of Alaska.

A warm episode also occurred in central Alaska, causing the glaciers to withdraw and permafrost with ice wedges to thaw. The ice wedges were replaced with windblown sand that filtered down from the overlying sand dunes forming ice wedge pseudomorphs (fig. 33). The lowering of the permafrost table and disappearance of the ice wedges indicate that the mean annual air temperature was at least warmer than 0°C. From both sea and land, then, comes strong evidence of a major warm interglacial period perhaps 1 m.y. ago.

MIDDLE PLEISTOCENE

In several parts of Alaska there is evidence of two or more Pleistocene cold periods of pre-Illinoian age. The exact stratigraphic positions are unknown, but one or two of the cold periods lie in what may roughly be called middle Pleistocene time. The Dry Creek Glaciation in the Alaska Range (Wahrhaftig, 1958) and the Caribou Hills Glaciation in Cook Inlet (Karlstrom, 1964) are not the earliest recorded glaciations in the respective areas, yet they are pre-Illinoian. In many other Alaskan localities, glacial deposits indicate a major cooling of the climate sometime in pre-Illinoian, perhaps in middle Pleistocene time (table 2). Pack ice, for example, appeared in the Arctic Ocean at least 700,000 years ago (Herman, 1970; Hunkins and others 1971).

In central Alaska, inactive solifluction deposits and altiplanation terraces record a major cold period. The upper of the two well-developed inactive solifluction layers on the campus of the University of Alaska (fig. 33) is probably middle Pleistocene in age. The inactive solifluction layer lies at an elevation of 150 m and is covered by loess of Illinoian and Wisconsinan age. Active solifluction occurs in central Alaska today at an elevation of 1,000 m.

Still preserved at elevations of 260–900 m are relict altiplanation terraces now covered by loess of Illinoian and Wisconsinan age. Well-developed but inactive altiplanation terraces of Wisconsinan age exist at 1,600 m near Fairbanks. A rigorous climate is necessary to form altiplanation terraces. Barring major tectonic movement, the low-level inactive solifluction deposits and relict altiplanation terraces at Fairbanks indicate that the climate was then much more rigorous than in any later Quaternary time (Péwé, 1970b).

The last of two or perhaps three periods of deposition of coarse, angular, local, auriferous gravel in creeks of central Alaska is middle Pleistocene in age (fig. 20). It is felt that the gravel is the debris shed from the hills by solifluction during the time that the ancient altiplanation terraces were cut (Péwé, 1970b) and, therefore, is additional evidence of a rigorous cold period.

In addition to the cold periods recorded in middle Pleistocene time, Hopkins (1967b) has evidence of an interglacial period, the Einahnuhtan transgression of between 250,000 and 50,000 years ago. The marine fauna indicates the temperature of the water was about the same as at present; no information is available on air temperature. According to Hopkins (1967b, p. 72), the transgression was preceded by a severe cold interval, because frost-disturbed beds under the Einahnuhtan beds on St. Paul Island indicate permafrost at low altitudes on the island. The warm interval cannot be directly associated with the cold intervals because of the lack of dating of the well-documented cold intervals of the middle Pleistocene in Alaska.

YARMOUTH(?)

Events associated with a climatic warming just prior to Illinoian time are treated here as Yarmouth(?) in age. Correlations with central United States and Europe time equivalents are probable, although not proved. The glacial record at various places in Alaska indicates that the immediate pre-Illinoian interglacial episode was an important warming period; however, more detailed information is available from western and central Alaska.

In the Bering and Chukchi Sea areas, deposits from the marine transgression of Yarmouth time (termed "Kotzebuan transgression" by Hopkins, 1965, 1967a) contain a molluscan fauna indicative of a water temperature the same as that now at Amchitka and St. Lawrence Islands. However, deposits of the Kotzebuan transgression around the shores of Kotzebue Sound and along the Siberian coast contain high-Arctic molluscan and foraminiferal faunas indicative of very cold water (Hopkins, 1972). McCulloch, Taylor, and Rubin (1965) referred to this transgression as Yarmouth in age near Kotzebue. Later, McCulloch (1967, p. 102) labeled the transgression as pre(?)-Illinoian on the basis of Th^{230}/U^{238} dates of 170,000 and 175,000 years (Blanchard, 1963). These dates may be suspect (Plafker, 1971) on the basis of the Th^{230}/U^{238} dates on shells by Blanchard from Middleton Island. Hopkins (1972) later stated that the marine beds at South Bight on Amchitka Island are of a Kotzebuan age, and uranium-thorium age determinations on bones and shells there are about 130,000 years old (Gard and others, 1972).

A study of pollen from the long core from Imuruk Lake (fig. 40) led Colinvaux (1964a, 1967a) to believe that the vegetation in central Seward Peninsula was similar to that now and that tree line was not too far distant. The summers were longer and warmer than those now.

From interior Alaska comes a record of a warmer climate, especially summers, as indicated by well-developed forests and thawing of permafrost and ice wedges. Underneath loess of Illinoian age in several exposures in the Fairbanks area is a well-developed forest bed, the Dawson Cut Formation (figs. 29, 41) (Péwé, 1952a, 1975).

It is thought that with warmer summers tree line rose and forests were widespread. Permafrost was gone, and loess deposition slowed, as suggested by geochemical analyses of silt of the forest bed (see p. 59; fig. 29). Thawing of permafrost indicates that the mean annual air temperature had warmed to at least 0°C or probably a few degrees warmer. Presence of ice wedge pseudomorphs in solifluction deposits at Fairbanks (fig. 33) and near Shaw Creek in central Alaska indicates warming of the climate either in Yarmouth time or earlier (Péwé, 1965b, fig. 4–22).

ILLINOIAN

EVIDENCE FROM GLACIATED AREAS

Extensive glacial advances during Illinoian time indicate a colder or wetter climate than now and of long duration. On the basis of cirque floors of Illinoian time, snowline was about 500–600 m lower than now and 150–250 m lower than its position during Wisconsinan time in the southwestern Seward Peninsula (Péwé and others, 1967). The rising of snowline to the east and north (figs. 11, 12) indicates the source of moisture was to the west and south and the climate became drier and more continental toward the interior of the State. R. Reger and V. Reger (unpub. data, 1971) calculated the mean July lapse rate for nearby Nome from 250 to 3,600 m elevation: 0.49°C per 100 m. On the basis of this vertical air temperature gradient and the 500-m lowering of snowline, the mean July air temperature in Illinoian time at Nome was 7.4°C, 2.4°C lower than today. By this method, mean July temperature at Indian Mountain near Hughes was probably at least 3.3°C lower than today. As previously noted, however, the amount of temperature drop calculated from cirque floor levels is a minimum figure.

EVIDENCE FROM UNGLACIATED AREAS

Abundant evidence indicates the presence of permafrost and ice wedges during Illinoian time in both central (Péwé, 1952a) and far western Alaska (McCulloch and others, 1965). These features reflect a rigorous periglacial climate with a mean annual air temperature of at least −7°C, 2°−4°C colder than the present mean annual air temperature in these localities.

Undoubtedly, periglacial frost features other than permafrost and ice wedges were present in unglaciated areas during Illinoian time, but as yet, none are recognized, except perhaps the rubble sheets on Jumbo Dome (Wahrhaftig, 1949). No quantitative data are available on the climate needed to produce the rubble sheets on Jumbo Dome; however, a rigorous climate must have existed.

Thick loess deposits of Illinoian age are widespread in central and western Alaska, indicating widespread glacial advances and extensive accompanying vegetation-free flood plains that existed a long time. The loess indicates great source areas and not necessarily higher winds or greater aridity. Silt deposits of Illinoian age in Fairbanks (fig. 20) have a conspicuous unconformity in the middle, which may mean a change in rate of deposition and perhaps in climate.

Biogeographical evidence to support the rigorous climate in Illinoian time is available but difficult to quantify. For the Imuruk Lake area, Colinvaux (1962) showed a colder climate than now, with an Arctic grass tundra and little or no alder and some dwarf birch; the tree line was not very close. In the Fairbanks area, tree line was lowered 600 m, and trees were sparse as the result of cooler and shorter summers (see p. 84).

In summary, the climatic conditions of at least central and western Alaska were rigorous and thought to be colder and of longer duration than during Wisconsinan time on the basis of more extensive glaciation and lower snowline. Ice wedge studies indicate that the mean annual air temperature in the central and western part of the State was at least −7°C, 2°–4°C colder than now. Snowline studies on the basis of cirque floor elevations suggest a depression of mean July temperature of 2.4°C in the far west. Reconstruction of forest distribution also indicates shorter and colder summers.

SANGAMON

During the Sangamon interval, the climate was warmer than now. Marine molluscan and fish fossils indicate that water temperatures off western and northern Alaska also were warmer in Sangamon time (Pelukian transgression of Hopkins) than at present (Hopkins, 1965). Plant remains interbedded with marine sediments indicate that the present forests of western Alaska extended 50–80 km more to the west, which suggests that the summers were appreciably warmer and longer than at present in that part of the State. This was also suggested by Colinvaux (1962) from his work at Imuruk Lake. He stated that the vegetation of Sangamon time was a sage-alder tundra similar to present conditions and tree line was nearby. McCulloch, Taylor, and Rubin (1965) also recorded westward advance of forests in Sangamon time into areas now tundra covered near Kotzebue. Hopkins (1972) cited R. L. Detterman to support the suggestion that spruce forest extended into the Arctic drainage of the Brooks Range in regions now supporting only shrub tundra.

The exposures near Fairbanks of white spruce-birch-aspen forest beds of Sangamon age indicate warm summers and severe winters. Summers were considerably warmer than in Illinoian and Wisconsinan time. However, vegetation alone does not reveal whether the mean summer air temperature was warmer than now.

From far western and central Alaska more quantitative paleoclimatic information is obtained from the study of past permafrost. The presence of ice wedge casts in now perennially frozen sediments of Illinoian age in both western (McCulloch and others, 1965; Hopkins, unpub. data) and central Alaska (Péwé, unpub. data, 1952) indicates that during the interval between Illinoian and Wisconsinan time the climate ameliorated to give a mean annual air temperature that was warmer than 0°C, thereby causing the ice wedges to melt.

The geochemical studies of perennially frozen sediments at Fairbanks (p. 59; fig. 29) (leaching of salts, as well as other evidence of ground-water movement through now perennially frozen sediments) further support the belief that most of or all the permafrost in central Alaska thawed in Sangamon time, and therefore, the mean annual air temperature was warmer than 0°C for a considerable length of time. Thawing of permafrost would necessarily thaw mammal carcasses preserved in the frozen ground; no carcasses of pre-Wisconsinan age are known, thereby strengthening the concept that permafrost perhaps thawed entirely in central Alaska in Sangamon time.

Formation of a weathering profile 3 m thick on the drift of the Nome River Glaciation (Illinoian in age) (p. 20) and on loess near Kotzebue (McCulloch and others, 1965) on the Seward Peninsula indicates that the permafrost table was lower or that permafrost was absent. On the north shore of Eschscholtz Bay during Sangamon time, there may have been a warm period followed by a cold period (regarded as pre-Wisconsinan in age) with formation of ice wedges, and then a return to a warm climate (Hopkins, 1965).

The climate of Sangamon time in central and western Alaska was warmer than now to allow permafrost to thaw and had longer and warmer summers to permit rise of tree line and further extension of it to the west and north. The mean July temperature in Sangamon time, for example, probably was warmer than the present mean July temperature of 9.9°C at Nome and 15.5°C at Fairbanks.

WISCONSINAN

The Wisconsinan Stage in Alaska represents the latest major cold phase in Quaternary time, as it does elsewhere; however, in this region work has not progressed in sufficient detail to divide climatic fluctuations of Wisconsinan time, except perhaps in a few areas. Therefore, all discussions of climatic conditions in Wisconsinan time will mean "Wisconsinan undifferentiated" unless otherwise indicated.

EVIDENCE FROM GLACIATED AREAS

A comparison of snowline maps of the present and Wisconsinan time for Alaska reveals quite strikingly the similarity of pattern of the isolines and of wind and moisture sources during these two times. This pattern and the close parallelism of modern and Wisconsinan snowlines strongly suggests that there probably was no increase in precipitation to produce Wisconsinan glaciers, but rather that glaciation probably was caused by a decrease in mean summer temperatures and increase in summer cloudiness which resulted in decreased snow and ice ablation.

Wisconsinan July mean temperatures were calculated from present lapse rate and the position of snowline in Wisconsinan time in many parts of the world (Charlesworth, 1957, p. 645). Values for Wisconsinan July mean temperatures range from 3° to 12°C. However, snowline was perhaps a bit lower than cirque floors (R. D. Reger and T. L. Péwé, unpub. data, 1974), especially for all but cirque glaciers. Nevertheless, elevation of cirque floors, an easily obtainable figure, is used throughout the world and therefore is used in this report. It may eventually be possible to calculate the position of snowline in Alaska from equilibrium lines (the level on a glacier where the net balance equals zero, and the accumulation equals ablation) on past glaciers. This should permit more accurate calculations of past temperature changes.

Calculations also assume that for a given locality the present lapse rate is the same as in Wisconsinan time. R. Reger and V. Reger (unpub. data, 1971) showed that this is probably not true. Especially in Alaska, proximity to the coast should probably be taken into consideration (R.D. Reger and T. L. Péwé, unpub. data, 1974). Table 14 shows that the lapse rate for coastal stations is much lower than for those in the interior part of the State. During Wisconsinan time, when the strait area was dry and stations such as Nome and Kotzebue were inland, the lapse rate was probably higher than now. However, the amount of difference is not known, and in this report modern lapse rate will be used; only minimum temperature changes are thus recorded.

A traverse from the west coast of Alaska to the interior reveals that snowline dropped more in the central part of the State than near the moisture-rich coasts (figs. 8, 9, 11, 12). Modern lapse rates also increase from coastal stations toward the interior. Therefore, the difference between Wisconsinan and modern mean July temperatures increases from the coast inland (table 14). In Wisconsinan time, the mean July air temperatures at Nome and Kotzebue were 8°C and 9.2°C, respectively, drops of 1.9°and 2.1°C. In the interior, however, Fairbanks and Northway show minimum drops of 4°C and 4.8°C for a mean July air temperature of 11.5°and 9.9°C respectively, in Wisconsinan times. At Indian Mountain near Hughes, the drop in mean July temperature was 2.7°C for early Wisconsinan time from calculations of past snowline on the basis of cirque floors. Calculations using the equilibrium line give a drop of mean July temperature of 3.9°C at Indian Mountain. Evidence of a middle Wisconsinan climatic change is the withdrawal of glaciers and transgression of the sea at this time.

TABLE 14.—*Present-day and Wisconsinan mean July temperatures of selected stations in Alaska calculated from present lapse rate and estimated change of snowline based on elevation of Wisconsinan cirque floors*

Station	Elevation (m)	Estimated lowering of snowline (m)[1]	Present July lapse rate per 100 m[2]	Elevation of interval of lapse rate (m)	Present July mean temperature[3] (°C)	Wisconsinan mean July temperature (°C)	Difference between present and Wisconsinan mean July temperature (°C)
Nome	5	400	0.49	250–3,600	9.9	8.0	−1.9
Kotzebue	3	400	.54	750–3,700	11.3	9.2	−2.1
Utopia	330	450	.61	500–3,600	[4]13.8	[5]11.1	[6]−2.7
Fairbanks	150	600	.67	500–3,600	15.5	11.5	−4.0
Northway	522	600	.81	1,250–3,700	[2]14.7	9.9	−4.8
McGrath	102	450	.62	250–3,700	[2]14.5	11.6	−2.9
Anchorage	38	300	.58	250–3,700	13.8	12.1	−1.74
Anaktuvuk Pass[7]	700	[8]700	.59	1,475–3,000	10.3	[9] 6.5	−3.8

[1]*Estimated* from figures 11, 12.
[2]From R. Reger and V. Reger (unpub. data, 1971).
[3]From U.S. Weather Bureau (1943).
[4]R. D. Reger and T. L. Péwé (unpub. data, 1974).
[5]Early Wisconsinan.
[6]A figure of 3.9°C was obtained by R. D. Reger and T. L. Péwé (unpub. data, 1974) calculating early Wisconsinan snowline from the equilibrium line of the past glacier.
[7]From Porter (1966, p. 93).
[8]Modern snowline placed above existing glaciers by Porter.
[9]Classical Wisconsinan.

EVIDENCE FROM UNGLACIATED AREAS

Permafrost was more extensive during Wisconsinan time than now. Deep-lying relics of Wisconsinan permafrost are present in the glacial outwash sediments near Big Delta (Péwé and Holmes, 1964; Péwé and others, 1969), as well as along the south border of the permafrost area in Alaska (fig. 22). Such information means only that the mean annual air temperature was −1°C or colder. Along the south border of the State this indicates a rise in the mean annual air temperature of a few degrees centigrade since Wisconsinan time.

In northern and northwestern Alaska, ice wedges were actively growing in Wisconsinan time and still are. All that can be deduced from this information is that the climate there in Wisconsinan time was at least as cold and snow cover as thin as now. In central Alaska, however, inactive ice wedges are widespread. If large, typical ice wedges such as are present require a mean annual air temperature of −6° to −8°C or colder to form, then the mean annual air temperature near Fairbanks, for example, would have to have been at least 3°−4°C colder than at present. The mean annual air temperature of the Fairbanks area thus must have been at least −7°C in Wisconsinan time.

Many of the large deep-lying ice wedges in the Fairbanks area have flat tops, indicative of downward thawing of permafrost and ice wedges. This thawing was dated at more than 38,000 years old (Sellmann, 1967) and is thought to represent a short middle Wisconsinan, or earlier, warm period with a mean annual air temperature of more than 0°C.

If the large-scale polygonal ground in the Bristol Bay area (Hopkins, Karlstrom, and others, 1955, pl. 39) is the result of ice wedge growth, as believed by the writer, the Wisconsinan mean annual air temperature in that area was at least −6° to −8°C, 8°–10°C colder than at present.

Preliminary results of work by R. D. Reger and T. L. Péwé (unpub. data, 1974) on paleoclimatic implications of altiplanation terraces indicate that at an elevation of 840 m on Indian Mountain near Hughes in western Alaska, the mean July temperature was at least as cold as about 5° or 6°C when the terraces were forming in Wisconsinan time. By extrapolation, it was at least that cold in July in the highlands of the Yukon-Tanana Upland, in the Amphitheater Mountains 30 miles west of Paxson, in the highlands of the Seward Peninsula, and elsewhere where altiplanation terraces formed during Wisconsinan time.

Other periglacial frost features, such as solifluction deposits, rubble sheets, rock glaciers, and patterned ground, are present in Alaska, both in the active and inactive form, but are less quantitative indicators of Wisconsinan climatic changes than ice wedges, altiplanation terraces, or permafrost. Throughout central and western Alaska a colder climate in the Wisconsinan glacial phase is indicated by the lower altitudinal limit of solifluction (fig. 8) (T. L. Péwé, unpub. data, 1964), rubble sheets (Wahrhaftig, 1949; Péwé, 1961c), and other frost features at that time.

There has been little investigation of paleowinds, or winds of Wisconsinan time, in Alaska similar to studies that have been done in Europe. Except for one or two local references, no critical work has been done, and in fact some misinterpretations of the paleoenvironment of the eolian sand and loess distribution are known. To produce the widespread, now stabilized sand dunes, it was necessary either to have had more widespread and more powerful winds than exist now or to create a much more widespread source of sand and silt. To produce a greater source, it is necessary to remove much of the vegetation. This can be done by creating more arid conditions, by dropping the water table, or by greatly increasing the size and number of wide, braided flood plains and outwash fans of glacial streams. Withdrawal of the sea from shelving shores would create larger sandy beaches than now exist. Both the increased size and number of braided streams as well as the lowering of the sea level took place as glaciation occurred in Wisconsinan time in Alaska. For example, Hopkins (1972) stated that the sheet of dune sand and loess covering some 12,000 km² of northwestern Seward Peninsula is evidently derived largely from a blanket of marine sand left exposed beneath the present-day Chukchi Sea.

Until detailed work reveals new facts, the model the writer suggests for paleowinds is that the extensive eolian sand and loess of Wisconsinan age do not indicate former greater aridity of any substantial change in amount or direction of winds but instead indicate an extensive increase in poorly vegetated or treeless gravelly-braided flood plains and outwash fans and plains throughout Alaska.

Comparison of modern and Wisconsinan snowline maps (figs. 11, 12) suggests that there was no increase in precipitation in Wisconsinan time and that the major wind directions and sources of moisture were probably the same. Grass and some sage (Artemisia) perhaps were abundant on the well-drained, more extensive active flood plains, outwash fans, and outwash plains. These broad generally flat areas were favored by great herds of grazers such as the bison, mammoth, and horse. The writer believes that grazers, grasslands, and xeric vegetation can be explained by the great increase of the treeless flood plains, outwash plains, and active sand dune areas in Wisconsinan time, and it is not necessary to call on greater general aridity throughout the entire unglaciated part of the State. The rolling lower hillslopes and small creek bottoms were probably boggy, perhaps more so than today, because the permafrost level was higher in Wisconsinan time and drainage was even poorer than now. Peat layers are common in the frozen retransported loess in valley bottoms, contrary to suggestions by Hopkins (1972).

In the pioneer study of the oriented lakes of the North Slope of Alaska, Black and Barksdale (1949) suggested that the lakes were oriented by winds blowing northwest (or southeast), parallel to the long axis of the lakes. This direction is 90° from the present northeast winds and from the wind direction associated with the modern and Wisconsinan sand dunes. Later studies by Carson and Hussey (1959, 1960a, b, 1962, 1963) showed that the lakes were, and still are, oriented by winds from the northeast.

BIOGEOGRAPHICAL EVIDENCE

It has been shown that the position of tree line in Wisconsinan time was 450–600 m lower in central Alaska and had withdrawn eastward of its present position on the Seward Peninsula. Tundra replaced the forests. Such changes indicate a more rigorous climate in Wisconsinan time than now. But inasmuch as the forest boundary is probably independent of the severity of winter temperatures, such shifts in tree line indicate changes in summer climate. Summers were shorter and cooler than now. Hopkins (1959a) showed that the spruce forest in Alaska is restricted to areas that have at least 90 "degree days" per year when the temperature reaches or exceeds 10°C. "Degree days" above 10°C was calculated by multiplying the amount by which the mean monthly temperature of the warmer months exceeds 10°C by the number of days in these warm months. In Fairbanks, for example, today the mean annual number of "degree days" above 10°C is 440 (Hopkins, 1959a, table 1). During much of the Wisconsinan interval, at least, the hills in the Fairbanks area were forest free; therefore, it is assumed that the number of mean annual degree days above 10°C was less than 90. There were a few spruce in the valleys (fig. 42), and the valley bottoms probably were at the threshold value of 90 "degree days" above 10°C; thus, Fairbanks probably had a summer climate similar to Moses Point or Council on the Seward Peninsula today.

Terrestrial mammal remains also provide information about Pleistocene climates. As shown, all forms known from Quaternary deposits could have lived under severe conditions; many live today in central Alaska. Forms such as the bison, mammoth, and horse were grazers and thus indicate more widespread grasslands. Mountain sheep, ground squirrels, and lemmings live today in tundra above tree line. These species indicate the reduction of forest and the concurrent shorter, colder summers. The Wisconsinan frozen carcasses suggest that permafrost has existed since the time of the animal's death. The mean annual air temperature must therefore have been colder than −1°C since their death, except for short warm periods which did not greatly affect the permafrost. Further, the presence of certain ground beetles (Matthews, 1968b) in silts near Fairbanks indicates a tundra environment and colder, shorter summers in Wisconsinan time.

SUMMARY

Many lines of evidence converge to indicate that summers were colder and cloudier in Wisconsinan time. Lower snowline suggests that mean July temperature dropped by 3° or 4°C, and the distribution of tree line, mammals, and beetles suggests colder summers also. Permafrost, especially ice wedges, demonstrates that the mean annual air temperature was several degrees colder in many parts of the State. A mean annual air temperature of at least −6° to −8°C is indicated for central Alaska and perhaps even as far south as Bristol Bay.

Although active sand dunes and loess deposits were widespread in Wisconsinan time, there is no good evidence to prove higher wind velocities or greater aridity.

HOLOCENE

With the amelioration of climate since Wisconsinan time, the massive glaciers have withdrawn, and snowline has risen. The climatic record in Alaska does not illustrate a simple warming since Wisconsinan time but an oscillation of climate. Throughout the State, the rather abrupt warming that began 10,000 years ago is reflected in a dramatic change in vegetation, permafrost, and glaciers. In some areas the Holocene warming peaked in a thermal maximum (hypsithermal interval) from about 7,500 to 3,500 years ago; in others, the warming maximum was recorded earlier. The thermal maximum of 7,500–3,500 years ago is not only recorded as a time of glacial minimum, and thus a warm period, but also is well inscribed in the botanical record. In the Brooks Range and near Umiat on the north side of the range, the slight warming at about 6,000 years ago may have been responsible for an alder maximum recorded (Livingstone, 1955, 1957). The warming is further supported by a date of 5,600±230 years from frozen alder remains in soils north of the Brooks Range (Tedrow and Walton, 1964). To Tedrow and Walton, the fossils indicate that the climate was warmer than now; they believe that alder does not grow in the area today. However, R. L. Detterman reported (written commun., Dec. 10, 1970) that there are good stands of alder today in the Colville River valley near Umiat. Perhaps because of an incomplete record, no thermal maximum is documented in the pollen record from Barrow.

On the Alaskan Peninsula (Heusser, 1963a) and in south-central and southeastern Alaska, detailed work by Heusser (1960) indicates a cool moist period from about 10,000 to 8,000 years ago, followed by a warm and drier thermal maximum period of 8,000–3,500 years ago (table 3). On St. Lawrence Island (Colinvaux, 1967b) and at the MacKenzie River delta (Ritchie and Hare, 1971), pollen studies indicate that the thermal maximum peaked at about 5,000 years B.P.

A detailed glacial record of Anaktuvuk Pass in the Brooks Range suggests that a period of maximum warmth somewhere between 6,260 and 2,800 years ago, but no quantitative data as to the degree of warming are available (Porter, 1966, fig. 22).

In parts of Alaska the classic postglacial thermal maximum (hypsithermal interval) is not recognized. Instead, an earlier dramatic warming and vegetation change are apparent, or the vegetation record shows only gradual warming and forest establishment to the present. In coastal Alaska, from Nome northeastward to at least the Sagavanirktok River, a warming is recorded 10,000–8,000 years ago, during which the climate became milder than at any other time since the last interglacial interval (Hopkins, 1972). Summers were considerably warmer, but winters may have been as severe as now.

Beaver-gnawed wood 9,400–8,300 years old in now treeless areas (Hopkins and Giddings, 1953; Hopkins, 1963) and evidence of thawing of 3 m of permafrost and melting of ice wedges 10,000–9,000 years ago on the Seward Peninsula (Hopkins and others, 1960, p. 54–55) are examples of this warming. Hopkins (1972) found wood of a large willow 8,360 years old in an ice wedge cast near Bering Strait where only depauperate herbaceous tundra now exists.

Evidence from the Kotzebue Sound area, combined with stratigraphic studies and radiocarbon dating, thoroughly documents a well-developed warming between 10,000 and 8,300 years ago in western and northern Alaska (McCulloch and Hopkins, 1966). In addition, in the Barrow area, Douglas and Tedrow (1960) found a buried discontinuous organic horizon in permafrost that is between 8,200 and 10,000 years old (Brown, 1965a) and, as such, might reflect a warmer climate.

A recent discovery of a poplar log 8,400±300 years old (Detterman, 1970) in stream gravels near the Sagavanirktok River in northern Alaska some 50 km north of the present limit of poplar trees is thought to be further evidence of the early Holocene warm interval.

Not all studies in western Alaska support this early thermal maximum. Schweger's (1971) pollen study from Onion Portage (p. 10) representing the last 12,000 years reports no record of a thermal maximum but a gradual change from tundra vegetation to spruce cover as the climate ameliorated.

The warming of the climate to give a mean annual air temperature of 0°C or warmer has long been recognized in central Alaska from the lowering of the permafrost table and melting of tops of ice wedges. This conspicuous feature was noticed by Taber (1943) but attributed to the Yarmouth interglacial warming.

From stratigraphic studies in the forties, it soon became apparent that this warming with subsequent lowering of ice wedge tops and the permafrost table was a post-Wisconsinan feature (Péwé, 1952a, 1958b, 1965a), possibly the hypsithermal interval. As more radiocarbon dates became available it was established that the warming began 10,000 years ago and the forest expanded from valley bottoms. The summers were considerably warmer.

Pollen studies from Fairbanks by J. V. Matthews (unpub. data, 1966) and from the Alaska-Canada border (Rampton, 1971) indicate only a general warming from 10,000 years ago but no hypsithermal warming.

The cause of the early thermal maximum is not clear.

McCulloch and Hopkins (1966) suggested that the warming in the coastal areas may have been due to a lower sea level and more widely exposed continental shelf from 10,000 to 8,000 years ago. However, Hopkins (1972) stated that new work by C. H. Nelson (unpub. data, 1970) indicates sea level was not low enough to expose a wide continental shelf. Also it is not clear how this would seriously affect interior Alaska. At the present time, the cause of the early thermal warming is undetermined.

Since the early or middle Holocene thermal maximum, glaciers have advanced in most parts of Alaska. Advances have been most extensive in south-central Alaska (Plafker and Miller, 1958) and indicate a cooler or moister climate, as also indicated by Heusser's study (1960) of the pollen record in the same area. A cooler climate since 3,500 years ago is also recorded in the Brooks Range (Livingstone, 1955, 1957). In the Fairbanks area, permafrost has reformed in the sediments that were thawed starting 10,000 years ago, indicating a return to a mean annual air temperature colder than 0°C. However, the climate is not cold enough or snowfall is too great to permit ice wedge growth, except in rare localities.

Rock glaciers form only in a permafrost climate and at altitudes lower than snowline; therefore, they are excellent indicators of these parameters. Wahrhaftig and Cox (1959) indicated that the many rock glaciers of the Alaska Range formed after the hypsithermal interval. This indicates that the mean annual air temperature cooled until the interstitial water remained as ice and, therefore, a mean annual air temperature colder than 0°C was reached. Yet, this cooling was not sufficient to bring the snowline down low enough to create new glaciers in these valleys. Therefore, the mean July temperature, although colder than in the thermal maximum, did not drop the 3°–5°C necessary to produce glaciers as it did in Wisconsinan time. Wahrhaftig and Cox believed that the two relatively cold periods that gave rise to corresponding periods of rock glacier formation consisted of a 1,000–2,000 year cold period about 4,000–2,000 years ago and another cold period of perhaps 600–1,000 years within the last millennium.

Tree-ring studies in southeastern Alaska (Lawrence, 1950) and interior Alaska (Péwé, 1951b, 1957b) and lichenometry (Reger and Péwé, 1969) suggest that the greatest extent of glaciers since the thermal maximum was reached about 200–300 years ago and that the climate has since become warmer. Several photographic sequences covering almost 100 years are available, showing glacier retreat and thinning in the Brooks Range (Hamilton, 1965b), central Alaska (T. L. Péwé, unpub. data, 1952), and southeastern Alaska (Field, 1932b, 1942, 1947; Cooper, 1937). Maps covering glacier retreat for about the last 200 years are available for areas in southeastern Alaska.

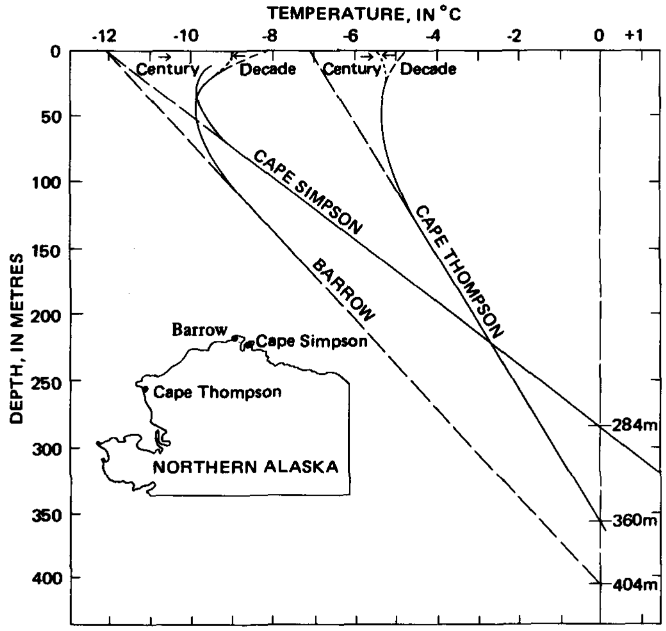


FIGURE 45.—Temperatures measured at three locations in Arctic Alaska (solid lines). Extrapolations are shown as broken lines. From Lachenbruch and Marshall (1969).

Quantitative data on climatic changes in northern Alaska over the last 100 years are available. Lachenbruch and Brewer's (1962) study of thermal records of deep permafrost near Barrow (fig. 45) indicates that the mean annual ground surface temperature rose about 4°C since about 1850, with about half of the increase occurring since 1930. In the Ogotoruk Creek area of northwestern Alaska, the thermal record of a borehole 357 m deep in permafrost approximately 1,200 m from the Chukchi Sea indicates that the mean ground surface temperature rose by 2°–2½°C in the last 100 years (Lachenbruch and others, 1966). Lachenbruch showed that the mean annual surface temperature at Barrow 100 years ago was about −12°C, and with the recent warming it should be about −8°C. However, the mean annual surface temperature today is slightly colder than −9°C (Lachenbruch and others, 1962). He demonstrated that recent cooling must have taken place and has penetrated to a depth of only 30 m (fig. 45). This recent cooling is also recorded in the thermal profile from Cape Thompson (Ogotoruk Creek). Such a cooling probably has been in progress only about 10 years (Lachenbruch and Marshall, 1969).

This warming of about the last 100 years is further substantiated by Hamilton (1965a), who deduced from a study of U.S. Weather Bureau records from the late

1800's to the present that there has been a net gain in air temperature of about 0.5°C–0.75°C.

Moore and Giddings (1962) suggested that wind direction for the past 5,000 years can be deduced from beach ridges at Cape Krusenstern. They hypothesized that the wind direction changed periodically as the polar front shifted. Additional information from elsewhere in the Arctic should be sought to more fully evaluate this idea.

Thus, in far western, northern, and central Alaska, maximum postglacial warming was reached early in the Holocene with cooling later. A fairly well documented record exists outlining a warming of the climate from rigorous Wisconsinan time climaxing in the thermal maximum of 7,500–3,500 years ago in some parts of Alaska. During this time the permafrost table dropped, the vegetation patterns changed slightly, and snowline rose. Since the early or middle Holocene warming, glaciers have advanced, rock glaciers have formed, the permafrost table has risen, and the climate has cooled slightly overall. However, glacier retreat and thermal records from permafrost indicate a warming over the last 100 years.

LATE CENOZOIC HISTORY OF ALASKA

A record of many events over the last few millions of years has been reviewed in the preceding pages, events that form a geological, biogeographical, and climatological history of Alaska during the development of modern landscapes, vegetation patterns, and faunal distributions, and the coming of man. Landscapes have been identified that, at one time, were forested, with glaciers mere pockets of ice in the high mountains; at another time landscapes were dominated by a tundra vegetation, disrupted by wide, braided glacial streams from which sand and dust were blown, and magnificient glaciers flowed from the ranges, spreading piedmont ice sheets on the lowlands. Other landscapes existed when sea level was lower, exposing wide tundra-covered shelves, then the home of large grazing land mammals; such terrains at other times were under the sea. Climates varied from those colder than now when ice wedges and permafrost existed even in southern Alaska and altiplanation terraces, solifluction lobes, and rubble sheets were actively forming at elevations hundred of metres lower than at present, to climates when permafrost was essentially absent in central Alaska and conditions were more equable than now in most of the State. A review of the late Cenozoic history can be pieced together from large and small fragments of information concerning the past 2–5 m.y. in Alaska; this history reveals controversies and paradoxes in a chronological review of events of late Cenozoic time and points up problems that remain to be solved.

LATE TERTIARY

Mountain glaciation probably began about 10–13 m.y. ago in the St. Elias Range and other ranges facing the Gulf of Alaska, and magnificent tidewater glaciers poured seaward. The earliest evidence of glaciation is about early middle Miocene (Miller, 1961b; Plafker, 1971); Bandy, Butler, and Wright (1969) believed it could be as young as late Miocene. The glaciers repeatedly formed ice shelves and discharged icebergs into the ocean (Taliaferro, 1932; Miller, 1953a, 1957; Plafker, 1971; Bandy and others, 1969; Kent and others, 1971; Scholl and others, 1971). In the Gulf of Alaska a mudstone accumulated at this time with abundant striated clasts and completely ungraded textures. The sediment consists of material that "rained" upon the sea bottom from floating glacier ice. This clastic mudstone is interpreted as glacial drift and is today part of the Yakataga Formation (table 1). Large glaciers also existed on the flanks of the Wrangell Mountains as early as 10 m.y. ago (Denton and Armstrong, 1969).

The extremely rugged and tectonically active glacier-clad terrain bordering the Gulf of Alaska in late Tertiary time was probably one of high relief. The climate was of the cool, temperate type such as exists today, not intensely cold, and rainfall was heavy. In fact, Hopkins (1972) mentioned that leaves of beech (*Fagus* sp.) were found with an early Pliocene mollusk fauna in the fine-grained beds interstratified with the tillites, indicating that land climates at sea level were somewhat milder than at present. The glaciers probably invaded a forest of beech trees.

The oceans off the southern coast into which the glaciers discharged were about 10° colder than now, as indicated by the cold-water planktonic foraminifer *Turborotalia pachyderma* (Bandy and others, 1969). Surface-water temperature decreased about 10°–15°C in late Miocene time, as indicated by molluscan and foraminiferal faunas (George Plafker, written commun., Apr. 21, 1970).

Little is known of glaciation elsewhere in Alaska prior to Quaternary time; however, evidence suggests (Clyde Wahrhaftig, oral commun., July 4, 1970) that one of the most extensive and earliest glaciations (the Browne Glaciation) on the north side of the Alaska Range may be 2.7 m.y. old. If so, this would indicate that the Alaska Range was a prominent topographic barrier to moisture-bearing winds from the south before the Pleistocene began (fig. 12).

In the highlands adjacent to the Arctic Basin, cooling and glaciation undoubtedly occurred 4–6 m.y. ago, as indicated by ice-rafted pebbles in the sediments on the sea floor (Steurwald and others, 1968). Although the waters of the Arctic Ocean must have begun to cool slightly, Hopkins (1972) pointed out that the Arctic Ocean probably was either ice free or only seasonally ice bound prior to 3 m.y. ago because the rich fauna of the North Pacific mollusks migrated to the Atlantic Ocean by way of Bering Strait about 3.5 m.y. ago and later, the time of the Beringian transgression.

Hopkins further stated that the recent discovery of *Turritella* in Pliocene beds offshore indicates temperate shelf waters during Pliocene time. Later, ice-rafting must have increased in the Arctic Ocean and eventually pack-ice cover formed (table 1). However, the Arctic Ocean probably remained essentially free of pack ice throughout the Pliocene epoch (Herman, 1970; Herman and others, 1971).

In general, the climate of late Tertiary time in Alaska must have been relatively equable and not intensely cold, although glaciers were present in the southern highlands during part of Miocene and all of Pliocene time. Little evidence is available concerning Alaskan vegetation in late Tertiary time, but in northern Seward Peninsula a forest of spruce, birch, pine, and hemlock existed 5.7 m.y. ago (Hopkins and others, 1971), indicating an environment similar to southeastern Alaska and coastal British Columbia rather than the Arctic climate of today. In general, the flora of the time of the Beringian transgression indicates that western and northern Alaska was considerably warmer than at present.

In late Tertiary time the Bering land bridge existed at various intervals. The Pacific and Arctic Oceans became connected by a shallow, sinuous seaway in late Miocene time, but the land bridge was reestablished about 5 m.y. ago. However, 3.5 m.y. ago, during the time of the Beringian transgression, there was free marine communication between the Bering and the Chukchi Sea.

Apparently toward the end of Pliocene time, the climate began to become more rigorous when perhaps the glaciers grew and the climate cooled to produce frost features. Hopkins (1972) reported boulder pavements created by frost action on basaltic lavas as old as 2 m.y. on St. George Island. Also, in very late Pliocene or very early Pleistocene time in central Alaska, angular, coarse auriferous gravel was being created in part by cutting of altiplanation terraces by nivation. The frost rubble produced by nivation and hillside rubble produced by frost action and other weathering processes moved downhill by solifluction to be buried by later Quaternary sediments. The tundra cover and associated colder climates probably did not arise in northern Seward Peninsula until early Pleistocene time.

EARLY PLEISTOCENE

The early Pleistocene history of Alaska is less clearly seen than the late Tertiary or the later part of the Pleistocene. Massive glaciers drained from mountains on the north side of the Gulf of Alaska and discharged into the sea, forming marine till. The fauna in most of the marine till indicates that part of the sediment is equivalent in age to the Anvilian transgression and later. In fact, sediments in the upper part of the section described by Miller (1953a) from Middletown Island represent the earliest Pleistocene ice advances and represent continual deposition from Pliocene time. Also, large glaciers of early Pleistocene age existed in parts of the St. Elias Range and Wrangell Mountains (Denton and Armstrong, 1969). Large glaciers evidently poured south from the Alaska Range (pl. 1) through the Cook Inlet trough and probably represent the Mount Susitna Glaciation of Karlstrom (1964). On the Seward Peninsula, glaciers moved south from the Bendeleben Mountains and entered the sea. This glaciation, called the Iron Creek Glaciation by Hopkins (1972), is thought to be 1.0±0.5 m.y. old. The glacial deposits underlie beds of the Anvilian transgression (table 8) that are less than 1 m.y. old but more than 700,000 years old.

Little information is available regarding temperature of the sea water in early Pleistocene time, but mollusks, particularly *Portlandia,* dredged from rocks of early Pleistocene age in the southern Bering Sea indicate that water temperatures were colder than now (Hopkins and others, 1969). *Portlandia* is not known to occur south of the Bering Strait at the present time.

Undoubtedly, the ice cover on the Arctic Ocean was becoming more extensive than in late Pliocene time, because by about 1 m.y. or 700,000 years ago, the permanent icepack was established (Herman, 1970; Herman and others, 1971). Steurwald, Clark, and Andrews (1968) believed that the Arctic Ocean was no warmer than at present for as long as the last 1.5 m.y.

The cooling of Alaska in early Pleistocene time is further borne out by periglacial phenomena. Ice wedge pseudomorphs are reported at Cape Deceit on the south shore of Kotzebue Sound (Guthrie and Matthews, 1971), and Hopkins (1972) believed they represent a permafrost period of 1.0±0.5 m.y. ago.

Around the beginning of Pleistocene time, the valleys of the small streams and rivers in central Alaska began to receive great amounts of angular, coarse auriferous gravel as altiplanation terraces were cut on the glacier-free highlands in the periglacial environment. The angular gravel was carried downhill by solifluction to accumulate in the valley bottoms, inasmuch as most of the material could not be removed by the drainage of the time. Remnants of solifluction layers of this age are widespread in central Alaska and contain well-developed ice wedge pseudomorphs (fig. 33). The ice wedges are thought to be at least 1 m.y. old because they represent the older of two permafrost periods, both of which predate the Illinoian Glaciation. Central Alaska must have had a mean annual air temperature less than −6° to −8°C at this time.

On the north side of eastern Seward Peninsula during a very early glaciation, thought to be correlative with the Iron Creek Glaciation, the forests were absent, and a more severe tundra than now existed. At this time nonwoody peat formed. However, just prior to this cold period and in very early Pleistocene time, a forest tundra existed, and the tree line advanced as far as Cape Deceit (Guthrie and Matthews, 1971); the vegetation was vastly different from that of late Tertiary time.

There is a clear record of repeated dispersals of land mammals from Asia to Alaska through much late Cenozoic time. The environment was such that most northern species of mammals could migrate across the land bridge; however, animals that inhabited forested mountain regions found it difficult or impossible to move across the open tundra landscape. The musk deer as well as the woolly rhinoceros were confined to the forest areas and did not move from Asia into Alaska. Many genera and species of animals moved across the bridge into Alaska much earlier than ordinarily recognized. In the Cape Deceit Formation (Guthrie and Matthews, 1971), there is evidence that in rather early Pleistocene time *Rangifer, Equus,* and *Microtus* had already moved into Alaska as had *Pliomys.* Other mammals that definitely were in Alaska in the early Pleistocene were *Canis, Cervus, Ochotona, Lemmus,* as well as an early collared lemming, *Predicrostonyx hopkinsi* (table 11).

MIDDLE PLEISTOCENE

The nature of the subaerial climate and landscape in Alaska is still poorly known in the rather ill-defined period of time from the earliest glaciation in the Pleistocene Epoch to the definite interglacial period prior to the Illinoian Glaciation. Nevertheless, enough significant data are becoming available from various areas to attempt to construct an outline of events during this interval.

After the early Pleistocene cold period, or periods, the Anvilian transgression occurred. This higher stand of sea level probably took place between 1.8 and 0.7 m.y. ago (Hopkins, 1972) (tables 1, 3). Warmer water temperatures than in the modern Bering and Chukchi Seas as well as the Arctic Ocean existed in Anvilian time, according to mollusk distribution. *Natica janthostoma* lived in the vicinity of Nome and Skull Cliff in northern Alaska but is now limited to waters adjoining Japan. During the warmer part of Anvilian time, winter ice probably did not reach Nome and may not have reached the Arctic coast of Alaska (Hopkins, 1967b). The Arctic Ocean was somewhat warmer than at present, but there is no unanimity as to whether pack ice was present (Herman, 1970; Herman and others, 1971; Hunkins and others, 1971).

A warm episode in central Alaska that possibly may be correlated with the Anvilian transgression in the west caused the glaciers to withdraw and permafrost to thaw. The ice wedges in solifluction deposits melted and were replaced with windblown sand that filtered down from the overlying sand dunes, forming ice wedge pseudomorphs (fig. 33). Lowering of the permafrost table and disappearance of ice wedges indicate that the mean annual air temperature was at least warmer than 0°C. From both marine and terrestrial deposits, then, comes strong evidence of a major warm interglacial period perhaps 1 m.y. ago.

In several parts of Alaska, there is evidence of two or more Pleistocene cold periods of pre-Illinoian age. The exact stratigraphic positions are unknown, but one or two of the cold periods lie in what may roughly be called middle Pleistocene time. After the warm period marked by the Anvilian transgression, the climate became more rigorous. Glaciers flowed northward into the Tanana valley from the Alaska Range during the Dry Creek Glaciation (Wahrhaftig, 1958) and Darling Creek Glaciation (Péwé and others, 1953). Larger glaciers pushed south from the ice-mantled Alaska Range and other ranges surrounding the Cook Inlet–Susitna valley trough to produce the massive trunk glacier of the Caribou Hills Glaciation (Karlstrom, 1964).

In other areas scant evidence suggests (table 2) that glaciers were quite extensive in this middle Pleistocene time in such areas as the Alaska Peninsula, the Brooks Range, and the Delta River area of the central Alaska Range. Undoubtedly, massive glaciers also continued to exist on the Wrangell Mountains and the ranges facing the Gulf of Alaska. Pack ice was probably widespread in the Arctic Ocean, since it appears to have been established at least 700,000 years ago (table 1).

Periglacial deposits in central Alaska record cold-climate processes active at this time. Solifluction lobes were moving downhill at an elevation as low as 150 m above sea level in central Alaska at this time. Today, active solufluction lobes exist only above an elevation of 1,000 m in central Alaska. Perhaps the upper of two well-developed inactive solufluction layers on the campus of the University of Alaska (fig. 33) is middle Pleistocene in age. Periglacial processes in middle Pleistocene time were active in the creation of altiplanation terraces from the Canadian border to far western and southwestern Alaska. Near the Fairbanks area, altiplantation terraces were being cut at elevations between 300 and 900 m above sea level, indicating a most rigorous climate because such terraces are not being cut today even at elevations of 1,600 m near Fairbanks.

Permafrost was widespread throughout Alaska during the one or more middle Pleistocene cold intervals, as indicated by the presence of ice wedges then. It is thought that permafrost existed even at low altitudes on St. Paul Island at this time (Hopkins, 1967b); permafrost does not exist on St. Paul Island today.

In addition to the cold periods recorded in middle Pleistocene time, an interglaciation is recorded by a high stand of sea level, the Einahnuhtan transgression of between 250,000 and 100,000 years ago (Hopkins, 1967b). The temperature of the sea water at that time was about as at present, and pack ice probably was present in the Arctic Ocean. The Bering Strait and the Arctic Ocean were connected at this time. The warm interval cannot be directly associated with the cold intervals because of the lack of dating of the well-documented cold periods of middle Pleistocene time in Alaska.

YARMOUTH(?)

The history of the period of climatic warming just prior to Illinoian time is clearer in Alaska than earlier Pleistocene events and is here treated as an interglacial episode, probably Yarmouth(?) in age. Although the correlations with the central United States and Europe time equivalents are not proved, evidence accumulated from marine deposits in the far west indicates that this is probably a correlative event.

During this time, the glaciers had greatly withdrawn, and sea level had risen probably 20 m above present sea level. This event was termed the Kotzebuan transgression by Hopkins (1965) and is correlative with the Holstein transgression of northern Europe. Hopkins (1972) thought it might correspond to the high sea level episode dated on many tropical shores at about 120,000 years old. Perhaps the marine beds at South Bight on Amchitka Island are of a Kotzebuan age; bones and shells there are about 130,000 years old (Gard and others, 1972).

During this high stand of the sea, the Pacific Ocean was connected to the Arctic Ocean. Hopkins (1972) believed that the usual northward circulation through Bering Strait was probably reversed at this time and cold water must have persistently flowed southward because high-Arctic molluscan and foraminiferal fauna around the shores of Kotzebue Sound, as well as on the Siberian side of the strait, indicate very cold water.

During the interglacial period, northwestern Seward Peninsula was forested, and spruce line advanced to at least as far as Cape Deceit. Today this area lies in the tundra region (fig. 40). In central Seward Peninsula summers were longer and warmer than now. In interior Alaska the climate was warm, in fact warmer than now, tree line rose, and forests were widespread. Well-developed forests were present, as indicated by the widespread forest beds at the base of the Illinoian loess (figs. 29, 41).

The thawing of the permafrost and melting of ice wedges in central Alaska indicate that the mean annual air temperature had warmed to at least 0°C and probably a few degrees warmer.

ILLINOIAN

Illinoian time in Alaska was one of maximum glaciation. Glaciers were extensive, and about 50 per cent of the State was covered by ice (fig. 6) the earlier glaciations probably were not much more extensive. The mountains in southern Alaska were clothed with massive valley glaciers and small icecaps which coalesced to form large trunk or trough glaciers as well as piedmont ice sheets. Ice from the Alaska Range moved south, completely blanketing the lowlands between the coastal ranges to the south; the ice then moved south and southwest toward the Alaska Peninsula. Ice was undoubtedly widespread in southeastern Alaska at this time, but no information is available. Local cirque and short valley glaciers were present in central Alaska. Ice coverage in Illinoian time in the Brooks Range was extensive (fig. 6) and much greater than in later Wisconsinan time, probably because most of the Brooks Range was high enough to intersect the Illinoian snowline but not high enough to intersect the higher Wisconsinan snowline. The major ice coverages in the Brooks Range and the Alaska Range–Coastal Range complex extended to the east, where they joined to form part of the Cordilleran ice sheet of North America.

Glaciers on Seward Peninsula pushed south from the central mountains and also were present on the far western tip of the peninsula. Glaciers from the Chukotka Peninsula of the U.S.S.R. adjacent to the Bering Strait pushed southward and eastward, reaching as far as St. Lawrence Island. Ice from U.S.S.R. probably pushed across dry land and did not exist as floating ice during the Kotzebuan transgression, as believed by Soviet workers (Hopkins, 1972).

Between the major ice-covered areas of the south and north in Alaska the Beringian refugium existed, which extended westward across the Bering land bridge. Many lobes of valley glaciers pushed into this refugium both from the Brooks Range on the north and the Alaska Range on the south (fig. 6). To the north of the Brooks Range an ice-free polar desert existed (Péwé, 1974a).

The altitudinal and areal distribution of the glaciers clearly indicates that as today, moisture sources were clearly to the south and the southwest in the north Pacific Ocean and the shrunken Bering Sea. Little moisture was contributed by air masses entering

Alaska from the Arctic Basin, as suggested by Ewing and Donn (1956, 1958). As in later time and today, the glaciers were more extensive in the southern part of the State than in the north and formed larger ice streams on the south side than on the north side of single mountain ranges (Karlstrom, 1964). Snowline was about 500–600 m lower than now and 150–250 m lower than its position during Wisconsinan time in different parts of Alaska. It was highest in the dry continental climate of east-central Alaska where it lay at an altitude of about 1,250 m during Illinoian time. Snowline sloped downward to an altitude of less than 300 m on Seward Peninsula in western Alaska (fig. 8).

If we visualize such an extensive glacial cover over Alaska in Illinoian time, we must, as a consequence, understand that all major streams and almost all minor streams drained glaciated terrain. Therefore, the glacial streams were undoubtedly choked with glacial silt, were aggrading, and had extensive braided, vegetation-free flood plains (fig. 42). Winds in the valleys blew silt from these barren flood plains and deposited it as loess over the countryside. Loess was deposited on ridges as high as 750 m above sea level, but most of it was deposited at lower altitudes, being thickest near streams. The Kuskokwim and Yukon Rivers wandered across the then dry floor of the Bering Sea, and from their flood plains, silt was blown to be deposited as loess on the Pribilof Islands. The greatest loess deposits are confined to central and western Alaska (fig. 17) and are locally as much as 50 m thick. Huge areas of dust blown from the flood plains were common throughout parts of Alaska, similar to areas of dust being deposited today as loess near such outwash streams as the Delta and Knik Rivers (pl. 1). Although sparse vegetation on flood plains is indicated by this wide distribution of eolian deposits, it does not necessarily indicate sparseness of vegetation away from the flood plains or aridity throughout Alaska, nor does it require exceptionally high winds.

Closer to the glaciers and mostly along the major streams, sand was blown, and widespread dune fields and sand blankets formed. Active dune fields were common along the Yukon, Tanana, Koyukuk, and Kuskokwim Rivers as well as near the flood plains of major rivers on the North Slope of Alaska.

The climate of the Illinoian glacial interval was rigorous, and abundant evidence indicates that permafrost was widespread. Ice wedges throughout most of Alaska reflect a periglacial climate with a mean annual air temperature of at least −7°C, about 2°–4°C colder than the present mean annual air temperature in central and western Alaska. Undoubtedly, periglacial frost features other than permafrost and ice wedges were present in unglaciated areas during Illinoian time, but as yet few have been recognized except perhaps rubble sheets on Jumbo Dome (Wahrhaftig, 1949).

In Illinoian time central Alaska and the North Slope were refugia, most probably with a tundra environment and perhaps with a scattering of some spruce and other trees in a few of the valley bottoms of central Alaska. The record of the flora in Alaska at this time is very scant and restricted to the Seward Peninsula and the Fairbanks area. In central Seward Peninsula, study of pollen from a core from Imuruk Lake (Colinvaux, 1962) indicates an Arctic grass tundra with little or no alder and some dwarf birch. Sage (*Artemisia*) was more common than now or in an interglacial period. Tree line was not close. On the northern part of Seward Peninsula at Cape Deceit, a tundra environment was present (Guthrie and Matthews, 1971).

Pollen analysis of samples of peat from Illinoian loess collected by the writer in valley bottoms of the Fairbanks area indicates that at this time tree line was significantly lower and trees were conspicuously rare. A lowering of tree line to creek valley bottoms required a drop of 450–600 m from today's tree line, suggesting that the climate, especially the summers, was indeed colder than now. A much more restricted forest area is also borne out by the greater number of grazers (especially mammoth, bison, and horse) then than now. The increase in *Artemisia* at this time probably does not support a case for widespread aridity in Alaska. *Artemisia* probably found a more favorable habitat on the almost vegetation-free sand dunes and flood plains of braided glacial streams that were so abundant at this time, rather than a general widespread increase of *Artemisia* throughout the entire tundra environment of hillslopes and boggy valley bottoms.

Although data are very meager, Hopkins (1972) believed that the exposed continental shelf of the Bering Sea at this time had a relatively maritime climate and supported a shrub tundra.

Over this widespread tundra environment wandered a large vertebrate population similar to that which existed in Wisconsinan time (fig. 42). Animals such as ground squirrels, lemmings, goats, and sheep limited to above or near tree line today lived at lower altitudes. The most abundant of all the mammals in Illinoian time were mammoth, horse, and bison. Despite information in modern literature to the contrary, *Rangifer, Praeovibos, Ovis, Alces, Ovibos, Saiga, Bootherium, Symbos,* and *Discrostonyx* as well as others were widespread in Illinoian time in Alaska. Perhaps most of them lived near the flood plain in winter where the winds blew much of the snow from the sparse vegetation; in summer they may have migrated to the lower slopes. *Saiga* is found nowhere else in North America but Alaska; this area represents the easternmost extension of its Siberian range, and southward penetration from Alaska was probably prevented by mountains and glaciers. *Saiga* is found both on the North Slope and in central Alaska during Illinoian time, and the environment favored by this animal is one that would have existed on the abundant windswept, flat, braided flood plains of glacial streams.

SANGAMON

The interglaciation between the Illinoian and Wisconsinan cold periods is well documented in Alaska. It was a time of great withdrawal of the glaciers, rising of sea level, thinning and disappearance of permafrost in many areas, and the return of the forests to most of the tundra regions, at least south of the Brooks Range.

With the warming of the climate in Sangamon time, the glaciers withdrew in Alaska as elsewhere around the world, and sea level rose. The rise of sea level and transgression onto the land in Alaska was termed the Pelukian transgression by Hopkins (1965). It is recorded widely throughout Alaska by well-preserved marine terraces and fairly well preserved marine scarps that occur 7–10 m above present sea level. Mollusks and fish present in the waters at this time off western and northwestern Alaska indicate that the sea was warmer than at present. For example, Hopkins (1972) mentioned that the deposits near Nome contain several species of mollusks such as *Pholadidea penita, Littorina littorea, Natica janthostoma,* and others which are now limited to more southern waters.

The warming of the climate is further recorded by the deep oxidation profiles present on Illinoian glacial moraines throughout central and western Alaska and by evidence of thawing of permafrost. Permafrost probably completely disappeared in Fairbanks, as indicated by the abundant ice wedge pseudomorphs, thick oxidation profiles in Illinoian loess, and absence of preserved mammal carcasses of Illinoian age. The mean annual air temperature undoubtedly was well above 0°C for a long period of time. Although it may have been absent in central Alaska, permafrost was still abundant in northern Alaska, and some was present in western Alaska, even though frozen ground melted to considerable depths and ice wedges melted (McCulloch, 1967).

At the time of thawing, the loess in central Alaska was deeply gullied, and block slumping occurred. Long parallel gulleys more than 10 m deep and 180 m long were cut in the easily eroded loess on slopes in the Fairbanks area.

The tundra environment present earlier in central Alaska was replaced by a spruce-birch forest. Forests were widespread, as indicated by buried forest layers dated at more than 56,900 years old between Illinoian and Wisconsinan loess (fig. 29). Summers were longer and warmer; the mean July temperature in the Sangamon, for example, was undoubtedly warmer than the present mean July temperature of 15.5°C at Fairbanks. The forest probably pushed northward to a position on the north side of the Brooks Range, as indicated by large logs found in the frozen ground (Hopkins, 1972), and advanced westward into the area of now treeless tundra (fig. 40) on the Baldwin Peninsula near Kotzebue (McCulloch and others, 1965). At Cape Deceit, however, a tundra environment existed; also, Colinvaux (1962) noted that at Imuruk Lake at this time the vegetation was a sage-alder tundra similar to the present conditions, but tree line was near. Colinvaux (1967b) suggested that in Sangamon time tundra was present on St. Lawrence Island, and all evidence so far indicates that the Alaskan side of the Bering Strait had a vegetation comparable to the present, even though spruce trees were 50–80 km west of their present limit in the Kotzebue Sound and Nome areas.

WISCONSINAN

In Wisconsinan time glaciers existed in nearly all the mountainous areas of Alaska. North of the crest of the Alaska Range and on the Alaska Peninsula, the glaciers were essentially the alpine type; they filled mountain valleys and in places spread as piedmont lobes in adjoining lowlands. However, central and northern Alaska were essentially ice free. Areas south of the crest of the Alaska Range and the Alaska Peninsula were almost completely inundated by ice, and icecap conditions prevailed over large areas (fig. 6). Glaciers filled adjoining lowland basins such as the broad Susitna valley and spread onto the then dry continental shelf, thickening to form icecaps as much as several tens of kilometres wide and several hundreds of kilometres long. The icecap in the Shelikof Strait sent glaciers northward where they penetrated passes into the mountainous backbone of the Alaskan Peninsula. Similar icecaps accumulated at several places in the Aleutian Range. During this latest major cold period, glaciers pushed south into Prince William Sound, and ice shelves were undoubtedly present. In the south half of Alaska the glaciers were almost as extensive as in Illinoian time. However, in northern Alaska the glaciers of the Brooks Range were much less extensive, as were those in the Bering Strait and on the Seward Peninsula (fig. 6). Wisconsinan glaciers may not have been very extensive along the Gulf of Alaska between Cape St. Elias and Russell Fjord because the mountains north of this part of the Gulf of Alaska coast were lower during Wisconsinan time.

Glaciers in the Copper River Basin did not extend far enough to completely cover this lowland, and ice blocked all the exits from the basin, producing a huge lake at least once during Wisconsinan time. The lake(s) existed for thousands of years, and laminated clays were interbedded with glacial deposits and volcanic mudflows. Undoubtedly, volcanic flows poured from the Wrangell Mountains into the Copper River Basin probably over glaciers or under glacial ice. Near the end of Wisconsinan time, drainage from the basin was reestablished, the lakes drained, and the former lake floor was entrenched.

In the western Brooks Range a large proglacial lake was formed by glacial blockage between the DeLong and Endicott Mountains.

A comparison between snowlines in Wisconsinan time and the present in Alaska reveals a strikingly similar pattern of isolines and of wind and moisture sources. The pattern and close parallelism of modern and Wisconsinan snowlines strongly suggest that increased precipitation did not produce Wisconsinan glaciers, but rather a decrease in mean summer temperatures and an increase in summer cloudiness, which resulted in decreased snow and ice ablation. Also, in Wisconsinan times the major sources of moisture, as in the past, were from the south and southwest in the Gulf of Alaska and the shrunken Bering Sea. There is no evidence that the ice-covered Arctic Ocean contributed enough moisture to sustain glacier growth. Interestingly, much of the precipitation for the south side of the central and eastern Alaska Range evidently entered through the Cook Inlet trough, as it does today. As indicated in figures 8 and 9, and especially 11 and 12, in Wisconsinan time there was an abrupt rise in snowline from near sea level in the Gulf of Alaska to about 1,500 m in interior Alaska. There was a steady rise from the Bering Sea on the west, eastward to Canada through central Alaska. In general, Wisconsinan snowline was about 300–400 m lower than modern snowline in the west and about 450–600 in the east, but local variations occurred. The vertical separation of the two snowlines becomes greater toward the interior because probably in Wisconsinan time, as now, the interior is drier than the far west.

Knowledge of the position of snowline in Wisconsinan time permits us to calculate past temperatures on the basis of the knowledge of the present lapse rate. For example, as indicated earlier, values for Wisconsinan mean July temperatures based on a critical evaluation of the present lapse rate and snowline in turn based on Wisconsinan cirque floors give us a broader view of summertime climate in central Alaska in Wisconsinan time (table 14). The difference between Wisconsinan and modern mean July temperatures is greater in the interior of Alaska than in the more cloudy, cool west. In Wisconsinan time, for example, the mean July air temperatures at Nome and Kotzebue were 8°C and 9.2°C, respectively, which is 1.9°C and 2.1°C lower than now. In the interior, however, at Fairbanks and Northway, mean July air temperatures were at least 4°C and 4.8°C lower in Wisconsinan time, if the temperatures there were 11.5°C and 9.9°C, respectively.

As in Illinoian time, all major and most minor streams drained glaciated areas and exhibited braided scantly vegetated flood plains which were excellent sources for silt and sand picked up and blown by the winds probably similar in velocity to those of today. Great quantities of sand were blown from these flood plains, glacial outwash fans, and outwash plains to produce extensive dune fields along the major streams such as the Tanana, Yukon, Koyukuk, and Coleville Rivers. These dunes, widespread as they were, were much more restricted than those of Illinoian time. With the lowering of sea level, large areas of the Bering Strait were exposed, presenting sand and silt to deflation. Hopkins (1972) believed that the great area of dune sand and loess covering 12,000 km^2 of northwestern Seward Peninsula was derived from a blanket of marine sand left exposed beneath the present-day Chukchi Sea. Also, the sand and loess on the Pribilof Islands date back to a time when the shoreline was much farther south than present.

During the Wisconsinan Glaciation, winds deposited a blanket of loess across central and western Alaska. Much of the loess deposited on the hills of central Alaska was retransported to valley bottoms to form an organic-rich, fetid, perennially frozen deposit. These deposits accumulated to a thickness as much as 60 m and incorporated abundant vertebrate and plant fossils of the time, including partial carcasses of vertebrates that did not completely decay prior to burial.

The gulleys and ridges of pre-Wisconsinan age in the loess on middle and upper slopes in the Fairbanks area were rounded and subdued by the Wisconsinan loess blanket.

The extensive eolian sand and loess of Wisconsinan age are interpreted not as an indication of former greater aridity or any substantial change in the amount and direction of winds but, instead, as in earlier major glaciations, as an extensive increase in poorly vegetated or nonvegetated sandy, gravelly, braided flood plains and outwash fans and plains throughout Alaska.

Permafrost in Alaska extended much farther south than at present (fig. 22), indicating that the mean annual air temperature was -1°C or colder at the south border of the State, a few degrees colder than now. In northern and northwestern Alaska, ice wedges were actively growing in the cold permafrost as they are today. In central Alaska permafrost and the mean annual air temperature may have been colder than today because ice wedges were actively growing then; today they are, for all practical purposes, inactive. Active ice wedges indicate that the mean annual air temperature was −6° to −8°C or colder; this would require a mean annual air temperature at least 3–4°C colder than at present. Perhaps it also indicates that the light, fluffy snow was somewhat more windblown and packed, probably because of the scarcity of trees.

Many of the existing large deep-lying ice wedges in the Fairbanks area have flat tops indicative of a downward thawing of permafrost and ice wedges sometime in Wisconsinan time. This thawing was dated as more than 38,000 years old (Sellmann, 1967) and is thought to represent a short middle Wisconsinan, or earlier, warm period when the mean annual air temperature was warmer than 0°C.

Rubble sheets, rock glaciers, and solifluction lobes were active on the slopes of central and western Alaska at lower altitudes than at present. Rubble sheets indicating a colder climate and more rigorous frost action were common on the south side (Péwé, 1961c) and the north side (Wahrhaftig, 1958) of the Alaska Range. Such frost-shattered and transported bedrock is also common in Seward Peninsula and even reported active on St. Paul Island almost at sea level during Wisconsinan time (Hopkins, 1972).

Approximately 50 percent of Alaska was ice free in Wisconsinan time and supported, for the most part, some form of tundra. The present taiga forest was greatly restricted and almost eliminated. Various types of tundra existed; in the far west Colinvaux reported (1967a) an Arctic herbaceous tundra, and Arctic sage was relatively common. Heusser (1963b) assumed that a tundra with little birch existed around Kotzebue Sound and Ogotorok Creek. On St. Lawrence Island was a tundra with grasses and sages predominating (Colinvaux, 1967a)

In central Alaska peat samples collected by the writer from Wisconsinan loess indicate that the valleys were covered with tundra rich in alder and dwarf birch with little grass. They also indicate that coniferous trees were scarce. In the valley bottoms at that time was a grass-sage-moss-*Equisetum* cover associated with small stands of dwarf birch and willow. The forest was not entirely absent as indicated by tree remains in beaver dams of late Wisconsinan age. Such dams include alder, spruce, willow, cottonwood, and birch wood.

Matthews (1970) reported that the spruce woodlands were nearly eliminated from interior Alaska in Wisconsinan time. Tree line in central Alaska dropped about 500 – 600 m, and the following tree-line mammals existed in lowlands near Fairbanks: ground squirrel, lemming, collared lemming, and mountain sheep. These animals now live at or above tree line on the tundra and not in the now-forested Fairbanks area.

During much of Wisconsinan time, grass and some sage (*Artemisia*) perhaps were abundant on the well-drained, much more extensive braided glacial flood plains, outwash fans, and outwash plains in unglaciated Alaska. These broad, flat areas were favored places for great herds of grazers such as the bison, mammoth, and horse, especially for part of the year. The habitat for grazers and the records of grasslands and xeric vegetation can most easily be explained by the great increase of these treeless flood plains and active sand dune areas in Wisconsinan time, rather than by a widespread arid condition throughout the unglaciated part of the State. The rolling, lower hill slopes and small creek bottoms were probably boggy, perhaps more so than today, because the permafrost level was higher in Wisconsinan time and drainage was even poorer than now. Peat layers were common in the valley bottoms.

It is also unnecessary to assume that a treeless vegetation prevailed in central Alaska in full glacial time and that the spruce persisted in a refugium somewhere in southwestern Alaska until about 6,000–8,000 years ago. Scattered spruce persisted in isolated creek bottoms in central Alaska (fig.42), inasmuch as spruce wood occurs in the frozen retransported silt of Wisconsinan age in the Fairbanks area. It seems more logical that repopulation of the forest of the interior and far west was from small nuclei preserved in the valleys of central Alaska. In fact, Young (1971) noted that the central part of Alaska would be expected to have a continental climate with summers warm enough to permit the growth of trees.

Throughout Wisconsinan time, the unglaciated refugia were occupied by an extensive mammal population, many species of which no longer exist today. As mentioned elsewhere, many more grazers were present then than now, and many of the tree-line and above tree-line species were more widespread than at present. The most abundant larger mammals were the mammoth, horse, and bison. It is in this environment that many of the carcasses of the animals were preserved because of the cold winters and short summers and eventually became buried to be perennially frozen until today.

Toward the end of Wisconsinan time, as the forest began returning, the ideal habitat of many of the grazers began to disappear. With the loss of the ideal habitat and increased predation by man, many mammals became extinct about 10,000 years ago. In late Wisconsinan time a marine transgression, the Woronzofian (Hopkins, 1967a), occurred, during which sea level was only a few metres lower than at present. Recent work by Schmoll, Szabo, Rubin, and Dobrovolny (1972) indicated that this transgression occurred around 14,000 years ago.

With the regressions and transgressions of sea in the Bering Strait area during Wisconsinan time, the Bering land bridge, the gateway from Asia to North America for many forms of life including man, was present for thousands of years and then narrowed and disappeared for intervals of time. It was demonstrated (Hopkins 1967a) that the bridge was present prior to 35,000 years ago but that it probably narrowed and was severed during middle Wisconsinan time, about 35,000–25,000 years ago. The land connection was restored during late Wisconsinan time, 25,000–12,000 years ago, although land connections may have been drowned occasionally for short periods during that interval. The land bridge was closed at the end of Wisconsinan time, between 11,000 and 10,000 years ago, when the sea rose.

HOLOCENE

About 10,000 years ago, the landscape began to take on aspects present today. Glaciers began to retreat, the forest began to return to much of the unglaciated terrain of central Alaska, and permafrost began to thaw and retreat northward. The record in Alaska does not illustrate a simple warming, retreat of the glaciers, or change of vegetation since Wisconsinan time but an oscillation of events. In general, the glaciers did begin to retreat about 10,000 years ago, but they still were in a rather advanced position on the north side of the Brooks Range, for example, from 8,300 to about 6,300 years ago (Porter, 1964). In most of Alaska the glaciers record the greatest retreat and farthest withdrawal at a time of thermal maximum, perhaps around 5,000–3,000 years ago. In southeastern Alaska the glaciers had withdrawn farther at that time than they have today (Goldthwait, 1963, 1966).

After this major post-Wisconsinan withdrawal, the glaciers advanced in the Brooks Range about 2,800 years ago. In southeastern Alaska they started to advance about 3,500 years ago, reaching a maximum probably only 300 or 400 years ago. In central Alaska it appears that greatest glacier advance after post-Wisconsinan thermal maximum was reached only as late as 200 or 300 years ago. In most parts of the State, glaciers have been retreating dramatically for the last 200 years.

In parts of Alaska the classic postglacial thermal maximum (hypsithermal) is not recognized. Instead, an earlier dramatic warming and vegetation change began around 10,000 years ago, and then the record shows gradual warming and forest establishment with some cooling to the present. In the west and northwest, the spruce forests were farther westward 9,400–8,300 years ago than now, and 9,000–10,000 years ago permafrost with ice wedges thawed down 3 m on the Seward Peninsula (Hopkins and others, 1960).

Around 10,000 years ago near Fairbanks, there was a thawing of permafrost, lowering of the tops of the ice wedges, and reintroduction of forest (Péwé, 1975). Subsequently, the permafrost table has risen. Pollen studies (Rampton, 1971) also indicate a major change of climate 10,000 years ago.

The vegetation changes in the Holocene in Alaska reflect many local differences. In some localities spruce appears to have returned before alder; in others, alder is the vanguard. A thermal maximum is recorded by vegetation changes in some areas and not in others. For example, at Onion Portage, Schweger (1971) reported no record of a thermal maximum but a rather gradual change from tundra vegetation to spruce cover. Further north, pollen work at Barrow does not demonstrate a thermal maximum, possibly because of an incomplete record. However, support for the advance of alder north of the Brooks Range in what might have been thermal maximum time is from alder leaves 5,650 years old preserved in the frozen ground in the Killik valley (Tedrow and Walton, 1964). Alder does not grow in the Killik valley today where these ancient leaves were recovered.

A record from Chandler Lake in the Brooks Range (Livingstone 1955, 1957) supports the presence of a thermal maximum and indicates that an herbaceous tundra was present 8,300 years ago, but at 7,500 years ago dwarf birch tundra became common. About 5,800 years ago, an alder maximum occurred which Livingstone thought was climatically related to the thermal maximum. To the south, on St. Paul Island of the Pribilof Islands, Colinvaux (1967c) noted that the pollen record shows that the vegetation similar to the present has existed for the last 9,000 years. In southeastern Alaska 10,000–8,000 years ago, pine was present (table 3), but from 8,000 to 3,500 years ago Sitka spruce and hemlock predominated (Heusser, 1960).

As in interglaciations before, the return of the forest to central and western Alaska was the most dramatic change in the vegetation in that part of the State. A controversy exists as to whether the exact time when it returned was 10,000 or 8,000 years ago. However, early in Holocene time the central and near western part of the State was indeed forested, and through Holocene time the forest boundary has evidently oscillated, especially on the north side of the Brooks Range and in the central and western Seward Peninsula. Hopkins (1972) believed that the Alaskan summers in the western part of the State are at present becoming warmer, and the greatest expansion of forests since the early Holocene warm interval appears to be happening now in northwestern Alaska.

Therefore, it is apparent that in some areas the vegetation type present today was essentially established within 1,000 or 2,000 years after the Holocene began, while in other areas, the modern vegetation dates back to a period immediately after the thermal maximum, only a few thousand years ago.

In addition to the advance of glaciers in the last 3,000 years, it should be noted that rock glaciers were active in the Alaska Range 4,000–2,000 years ago as well as during another cold period that occurred perhaps within the last 1,000 years (Wahrhaftig and Cox, 1959). The permafrost table rose in central and western Alaska since the warming of the early part of the Holocene. The advance of glaciers and rock glaciers and the rise in the permafrost table all point to a cooling of the climate in the last 3,000 years.

For the last 100 or 200 years, glaciers throughout Alaska have been rapidly retreating, and from a study of thermal records of deep permafrost near Barrow, it is evident that the annual ground surface temperature has increased about 4° since 1850 and half of this since 1930 (Lachenbruch and Brewer, 1962). Further warming was substantiated by examination of detailed weather records from the 1800's (Hamilton, 1965a).

REFERENCES CITED

Abbé, Cleveland, Jr., 1906, Climate, *in* Brooks, A. H., Geography and geology of Alaska: U.S. Geol. Survey Prof. Paper 45, p. 133–143.

Addicott, W. O., Kahno, Saburo, Sakamoto, Kenji, and Miller, D. J., 1970, Clark's Tertiary molluscan types from the Yakataga District, Gulf of Alaska, Alaska: U.S. Geol. Survey open-file report, 21 p.

Ager, T. A., 1972, Surficial geology and Quaternary history of the Healy Lake area, Alaska: Alaska Univ., M.S. thesis, 127 p.

Allan, R. J., 1969, Clay mineralogy and geochemistry of soils and sediments with permafrost in interior Alaska: Dartmouth College, Ph.D. dissertation, 289 p.

Alpha, T. R., 1970, Physiographic drawing of Nuka Bay, Alaska: U.S. Geol. Survey open-file map, 1 sheet.

American Geological Institute, 1972, Glossary of geology: Washington, D.C., 857 p.

Anderson, D. D., 1968, A Stone Age campsite at the gateway to America: Sci. American, v. 218, no. 6, p. 24–33.

——— 1970, AKMAK— An early archeological assemblage from Onion Portage, northwest Alaska: Acta Arctica, v. 16, 80 p.

Anderson, G. S., and Hussey, K. M., 1963, Preliminary investigation of thermokarst development on the North Slope, Alaska: Iowa Acad. Sci. Proc., v. 70, p. 306–320.

Anderson, L. A., and Johnson, G. R., 1970, Application of magnetic and electrical resistivity methods to placer investigations in the Fairbanks District, Alaska, *in* Geological Survey research, 1970: U.S. Geol. Survey Prof. Paper 700–C, p. C107–C113.

Anderson, S. T., and Bank, T. P., II, 1952, Pollen and radiocarbon studies of Aleutian soil samples: Science, v. 116, p. 84–86.

Andersson, J. G., 1906, Solifluction, a component of subaerial denudation: Jour. Geology, v. 14, p. 91–112.

Andreasen, G. E., Wahrhaftig, Clyde, and Zietz, Isidore, 1964, Aeromagnetic reconnaissance of the east-central Tanana Lowland, Alaska: U.S. Geol. Survey Geophys. Inv. Map GP–447, scale 1:125,000.

Andrews, J. T., 1965, The corries of the northern Nain-Okak section of Labrador: Geog. Bull., v. 7, no. 2, p. 129–136.

Andrews, J. T., and Miller, G. H., 1972, Quaternary history of northern Cumberland Peninsula, Baffin Island, N. W. T., Canada; Part 4, Maps of the present glaciation limits and lowest equilibrium line altitude for north and south Baffin Island: Arctic and Alpine Research, v. 4, no. 1, p. 45–59.

Anthony, H. E., 1949, Nature's deep freeze: Nat. History, v. 58, p. 296–301.

Auer, Väinö, 1920, Uber die Entstehung der Strange auf den Torfmooren: Acta Forestalia Fennica, v. 12, p. 23–145.

Bandy, O. L., Butler, E. A., and Wright, R. C., 1969, Alaskan upper Miocene marine glacial deposits and the *Turborotalia pachyderma* datum plane: Science, v. 166, p. 607–609.

Baranov, I. Ia., 1956, Map showing distribution of permafrost in the U.S.S.R., *in* Tsytovitch, N. A., 1958, Fundamentals and foundations on frozen ground: Akad. Nauk. SSSR Izv., 167 p. (in Russian).

Barnes, D. F., 1961, Gravity low at Minto Flats, Alaska, *in* Short papers in the geologic and hydrologic sciences: U.S. Geol. Survey Prof. Paper 424–D, p. D254–D258.

Barth, T. F. W., 1956, Geology and petrology of the Pribilof Islands, Alaska: U.S. Geol. Survey Bull. 1929–F, p. 101–160.

Bateman, A. M., and MacLaughlin, D. H., 1920, Geology of the ore deposits of Kennecott, Alaska: Econ. Geology, v. 15, p. 1–80.

Becker, R. E., and Volkmann, C. M., 1961, A preliminary report on the bacteriology of permafrost in the Fairbanks area: Proc. Alaskan Sci. Conf., College, Alaska, v. 12, p. 188.

Bell, E. K., 1974, Origin of the auriferous clays in the Fairbanks area, Alaska: Arizona State Univ., M.S. thesis, 63 p.

Bennett, H. H., and Rice, T. D., 1919, Soil reconnaissance in Alaska, with an estimate of the agricultural possibilities: U.S. Dept. Agriculture, Bur. Soils, Field Operations, 1914, 16th Rept., p. 43–236.

Benson, C. S., 1968, Glaciological studies on Mount Wrangell, Alaska, 1961: Arctic, v. 21, no. 3, p. 127–152.

Benson, Carl, Fahl, Charles, Trabant, Dennis, Well, Gunter, and Wendler, Gerd, 1970, Preliminary results of studies on the McCall Glacier, Brooks Range, Alaska: Univ. Alaska, Geophys. Inst., College, Alaska, 49 p.

Berger, A. R., 1960, On a recent volcanic ash deposit Yukon Territory: Geol. Assoc. Canada Proc., v. 12, p. 117–118.

Beskow, Gunnar, 1947, [Tjälbildningen och tjällyftningen] Soil freezing and frost heaving with special application to roads and railroads: Swedish Geol. Soc., ser. C, no. 375, 145 p. (English translation plus supplement by Osterberg, J. O., pub. by Northwestern Univ.)

Black, R. F., 1950, Permafrost, *in* Trask, P. D., Applied sedimentation: New York, John Wiley & Sons, p. 247–275.

——— 1951a, Eolian deposits of Alaska: Arctic, v. 4, p. 89–111.

——— 1951b, Structure in ice wedges of northern Alaska: Geol. Soc. America Bull., v. 62, p. 1423–1424.

——— 1952a, Polygonal patterns and ground conditions from aerial photographs: Photogramm. Eng., v. 18, p. 123–134.

——— 1952b, Structures in ice wedges of northern Alaska [abs.]: Am. Mineralogist, v. 37, p. 285–286.

——— 1954, Permafrost, a review: Geol. Soc. America Bull., v. 65, p. 839–855.

——— 1958, Lowlands and plains of interior and western Alaska, *in* Williams, Howel, ed., Landscapes of Alaska: Berkeley, Univ. California Press, p. 76–81.

——— 1963, Les coins de glace et le gel permanent dans le Nord de l'Alaska: Annales Géographie, v. 72, no. 391, p. 257–271.

——— 1964, Gubik formation of Quaternary age in northern Alaska: U.S. Geol. Survey Prof. Paper 302–C, 91 p.

——— 1966, Late Pleistocene to Recent history of Bering Sea—Alaska coast and man: Arctic Anthropology, v. 3, no. 2, p. 7–22.

——— 1969a, Geology, especially geomorphology, of northern Alaska: Arctic, v. 22, no. 3, p. 283–295.

——— 1969b, Thaw depressions and thaw lakes, a review: Biul. Peryglacjalny, v. 19, p. 131–150.

Black, R. F., and Barksdale, W. L., 1949, Oriented lakes of northern Alaska: Jour. Geology, v. 57, p. 105–118.

Black, R. F., and Laughlin, W. S., 1964, Anangula—a geologic interpretation of the oldest archeological site in the Aleutians: Science, v. 143, p. 1321–1322.

Blackwell, M. F., 1965, Surficial geology and geomorphology of the Harding Lake area, Big Delta quadrangle, Alaska: Alaska Univ., M.S. thesis, 91 p.

Blanchard, R. L., 1963, Uranium decay series disequilibrium in age determination of marine calcium carbonates: Washington Univ., Ph.D. thesis, 164 p.

Borns, H. W., Jr., and Goldthwait, R. P., 1966, Late-Pleistocene fluctuations of the Kaskawulsh Glacier, southwestern Yukon Territory, Canada: Am. Jour. Sci., v. 264, p. 600–619.

Bostock, H. S., 1952, Geology of northwest Shakwak Valley, Yukon Territory: Canada Geol. Survey Mem. 267, 54 p.

Boyd, W. L., and Boyd, J. W., 1964, The presence of bacteria in permafrost of the Alaskan Arctic: Canadian Jour. Microbiology, v. 10, no. 6, p. 917–919.

Bradley, C. C., 1948, Geologic notes on Adak Island and the Aleutian Chain, Alaska: Am. Jour. Sci., v. 246, p. 214–240.

Brewer, M. C., 1958, Some results of geothermal investigations of permafrost in northern Alaska: Am. Geophys. Union Trans., v. 39, p. 19–26.

Britton, M. E., 1958a, A tundra landscape (Alaska): U.S. Office Naval Research, Research Rev., p. 4–13.

——— 1958b, Vegetation of the arctic tundra, *in* Arctic biology: Oregon State Coll., Ann. Biology Colloquium, 18th, Corvallis 1957, p. 26–61.

Brooks, A. H., 1906, The geography and geology of Alaska, a summary of existing knowledge: U.S. Geol. Survey Prof. Paper 45, 327 p.

——— 1953, Blazing Alaska's trails: Caldwell, Idaho, Caxton Printers, 528 p.

Brown, Jerry, 1965a, Radiocarbon dating, Barrow, Alaska: Arctic, v. 18, p. 36–48.

——— 1965b, Seasonal thaw and ground-ice morphology in tundra soils, Barrow, Alaska [abs.]: Agronomy Abs., p. 59.

——— 1966a, Massive underground ice in northern regions: U.S. Army Conf. Proc., West Point, N.Y., v. 1, p. 89–102.

——— 1966b, Ice-wedge chemistry and related frozen ground processes, Barrow, Alaska, *in* Permafrost Internat. Conf., 2d, Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 94–98.

——— 1967a, Tundra soils formed over ice wedges, northern Alaska: Soil Sci. Soc. America Proc., v. 31, p. 686–691.

——— 1967b, Seasonal thaw chronology, Barrow, Alaska: U.S. Army Materiel Command Cold Regions Research Eng. Lab. Tech. Note, 4 p.

——— 1967c, An estimation of the volume of ground ice, coastal plain, northern Alaska: U.S. Army Materiel Command Cold Regions Research and Eng. Lab. Tech. Note, 22 p.

———1969a, Ionic concentration gradients in permafrost, Barrow, Alaska: U.S. Army Materiel Command Cold Regions Research and Eng. Lab. Research Rept. 272, 25 p.

———1969b, Soil properties developed on the complex tundra relief of northern Alaska: Biul. Peryglacjalny, no. 18, p. 153–167.

——— 1969c, Soils of the Okpilak River region, Alaska, *in* Péwé, T. L., ed., The periglacial environment, past and present: Montreal, McGill-Queens Univ. Press, p. 93–128.

——— 1969d, Buried soils associated with permafrost, *in* Pedology and Quaternary Research Symposium: Edmonton, Univ. Alberta Press, p. 115–127.

Brown, Jerry, Gray, Shelton, and Allan, Roderick, 1969, Evolution à la fin du Quaternaire d'un remblai de vallée à Fairbanks, Alaska, I; La géochimie et la stratigraphie du pergelisol [abs.]: Internat. Assoc. Quaternary Research, 8th Cong., Paris, Resumé des Commun., p. 270.

Brown, Jerry, Gray, Shelton, and Webster, William, 1967, Chemical and related properties of a permafrost section from Fairbanks, Alaska: U.S. Army Materiel Command Cold Regions Research and Eng. Lab. Tech. Note, 18 p.

Brown, Jerry, and Johnson, P. L., 1965, Pedoecological investigations, Barrow, Alaska: U.S. Army Materiel Command Cold Regions Research and Eng. Lab. Tech. Rept. 159, 32 p.

Brown, Jerry, and Sellmann, P. V., 1966a, Radiocarbon dating of coastal peat, Barrow, Alaska: Science, v. 153, p. 299–300.

Brown, Jerry, and Sellmann, P. V., 1966b, Pleistocene and Recent events in the Barrow, Alaska area: U.S. Army Materiel Command Cold Regions Research and Eng. Lab. Tech. Note, 3 p.

Brown, R. J. E., 1960, The distribution of permafrost and its relation to air temperature in Canada and the U.S.S.R.: Arctic, v. 13, p. 163–177.

——— 1966, Influence of vegetation on permafrost, *in* Permafrost Internat. Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 20–25.

——— 1969, Factors influencing discontinuous permafrost in Canada, *in* Péwé, T. L., ed., The periglacial environment, past and present: Montreal, McGill-Queens Univ. Press, p. 11–53.

———1970, Permafrost in Canada: Toronto, Canada, Univ. Toronto Press, 234 p.

Brown, R. J. E., and Péwé, T. L., 1973, Distribution of permafrost in North America and its relationship to the environment; 1963–73—a review, *in* Permafrost, The North American contribution to the Second International Conference, Yakutsk, Siberia, U.S.S.R.: Washington, D.C., Natl. Acad. Sci., p. 71–100.

Bryan, Kirk, 1946, Cryopedology—the study of frozen ground and intensive frost action with suggestions on nomenclature: Am. Jour. Sci., v. 244, p. 622–642.

Buckland, William, 1831, On the occurrence of the remains of elephants, and other quadrupeds, in the cliffs of frozen mud, in Eschscholtz Bay, within Bering's Strait, and in other distant parts of the shores of Arctic seas, *in* Beechey, F. W., Narrative of a voyage to the Pacific and Bering's Strait: London, Colburn and Bentley, p. 593–612.

Bull, Colin, 1969, The 1978–1980 surge of the Sherman Glacier, south-central Alaska: Canadian Jour. Earth Sci., v. 6, no. 4, p. 841–843.

Bull, Colin, and Marangunic, Cedomir, 1968, Glaciological effect of debris slide on Sherman Glacier, *in* The great Alaska earthquake of 1964 (Hydrology Volume): Natl. Acad. Sci. Pub. 1603, p. 309–317.

Bunge, Alexander von, 1884, Naturhistorische boebachtungen und fahrten im Lena-Delta: Acad. Imp. Sci. St. Petersbourg Bull., v. 29, p. 422–476.

——— 1902, Einige worte zur bodeneisfrage: Russian K. Min. Gesell. Verh., Sweite Ser., v. 40, p. 203–209.

Burns, J. J., 1964, Pingos in the Yukon-Kuskokwim Delta, Alaska; their plant succession and use by mink: Arctic, v. 17, p. 203–210.

Byers, F. M., Jr., 1959, Geology of Umnak and Bogoslof Islands, Aleutian Islands, Alaska: U.S. Geol. Survey Bull. 1028–L, p. 267–369.

Cairnes, D. D., 1912a, Differential erosion and equiplanation in portions of Yukon and Alaska: Geol. Soc. America Bull., v. 23, p. 338–348.

——— 1912b, Some suggested new physiographic terms: Am. Jour. Sci., v. 34, p. 75–87.
Campbell, J. M., ed., 1962, Prehistoric culture relations between the arctic and temperate zones of North America: Arctic Inst. North America Tech. Paper 11, 181 p.
Capps, S. R., 1910a, Rock glaciers in Alaska: Jour. Geology, v. 18, p. 359–375.
——— 1910b, Glaciation on the north side of the Wrangell Mountains, Alaska: Jour. Geology, v. 18, p. 33–57.
——— 1912, Glaciation of the Alaska range: Jour. Geology, v. 20, p. 415–437.
——— 1915a, Two glacial stages in Alaska: Jour. Geology, v. 23, p. 748–756.
——— 1915b, An estimate of the age of the last great glaciation in Alaska: Wash. Acad. Sci. Jour., v. 5, p. 108–115.
——— 1916a, An ancient volcanic eruption in the upper Yukon basin: U.S. Geol. Survey Prof. Paper 95–D, p. 59–64.
——— 1916b, The Chisana–White River district, Alaska: U.S. Geol. Survey Bull. 630, 130 p.
——— 1919, The Kantishna region, Alaska: U.S. Geol. Survey Bull. 687, 118 p.
——— 1931, Glaciation in Alaska: U.S. Geol Survey Prof. Paper 170–A, p. 1–8.
——— 1935, The southern Alaska Range: U.S. Geol. Survey Bull. 862, 101 p.
——— 1940, The geology of the Alaska Railroad region: U.S. Geol. Survey Bull. 907, 201 p.
Carsola, A. J., 1954, Extent of glaciation on the continental shelf in the Beaufort Sea: Am. Jour. Sci., v. 252, p. 366–371.
Carson, C. E., 1968, Radiocarbon dating of lacustrine strands in Arctic Alaska: Arctic, v. 21, no. 1, p. 12–26.
Carson, C. E., and Hussey, K. M., 1959, The multiple-working hypothesis as applied to Alaska's oriented lakes: Iowa Acad. Sci. Proc., v. 66, 334–349.
——— 1960a, Hydrodynamics in some arctic lakes: Iowa Acad. Sci. Proc., v. 67, p. 336–345.
——— 1960b, Hydrodynamics in three arctic lakes (Alaska): Jour. Geology, v. 68, p. 585–600.
——— 1962, The oriented lakes of arctic Alaska: Jour. Geology, v. 70, p. 417–439.
——— 1963, The oriented lakes of arctic Alaska, a reply: Jour. Geology, v. 71, p. 532–533.
Cass, L. A., and Miller, R. D., 1959, Role of the electric double layer in the mechanism of frost heavings: U.S. Army Corps Engineers Snow, Ice, and Permafrost Research Establishment, Research Rept. 49, 15 p.
Chaney, R. W., 1938, Paleoecological interpretations of Cenozoic plants in western North America: Bot. Rev., v. 9, p. 371–396.
Chaney, R. W., and Mason, H. L., 1936, A Pleistocene flora from Fairbanks, Alaska: Am. Mus. Novitates, v. 887, 17 p.
Chapman, R. M., Detterman, R. L., and Mangus, M. D., 1964, Geology of the Killik-Etivluk Rivers region, Alaska: U.S. Geol. Survey Prof. Paper 303–F, p. 325–407.
Chapman, R. M., and Sable, E. G., 1960, Geology of the Utukok-Corwin region, northwestern Alaska: U.S. Geol. Survey Prof. Paper 303–C, p. 47–164.
Chapman, R. M., Weber, F. R., and Taber, Bond, 1971, Preliminary geologic map of the Livengood quadrangle, Alaska: U.S. Geol. Survey open-file map, scale 1:250,000.
Charlesworth, J. K., 1957, The Quaternary Era: London, Edward Arnold, 1700 p.
Churchill, E. D., 1955, Phytosociological and environmental characteristics of some plant communities in the Umiat region of Alaska: Ecology, v. 36, p. 606–627.
Clark, D. L., 1969, Paleoecology and sedimentation in part of the Arctic Basin: Arctic, v. 22, no. 3, p. 233–245.
——— 1971, Arctic ice cover and its late Cenozoic history: Geol. Soc. America Bull., v. 82, p. 3313–3324.
Coats, R. R., 1956, Reconnaissance geology of some western Aleutian Islands, Alaska: U.S. Geol. Survey Bull. 1023–E, p. 83–100.
Cobb, E. H., and Kachadoorian, Reuben, 1961, Index of metallic and nonmetallic mineral deposits of Alaska, compiled from published reports of federal and state agencies through 1959: U.S. Geol. Survey Bull. 1139, 363 p.
Colbaugh, P. R., 1968, The environment of the Imuruk Lake area, Seward Peninsula, Alaska, during Wisconsin time: Ohio State Univ., M.S. thesis, 118 p.
Colinvaux, P. A., 1962, The environment of the Bering Land Bridge: Duke Univ., Ph.D. thesis, 260 p.
——— 1963, A pollen record from arctic Alaska reaching glacial and Bering land bridge times: Nature, v. 198, p. 609–610.
——— 1964a, The environment of the Bering Land Bridge: Ecol. Monograph, v. 34, p. 297–329.
——— 1964b, Origin of ice ages; pollen evidence from arctic Alaska: Science, v. 145, p. 707–708.
——— 1965, Vegetational history of arctic Alaska and islands of the Bering Sea: Science, v. 147, p. 633.
——— 1967a, Quaternary vegetational history of Arctic Alaska, *in* Hopkins, D. M., The Bering Land Bridge: Stanford, Stanford Univ. Press, p. 207–231.
——— 1967b, A long pollen record from St. Lawrence Island, Bering Sea (Alaska): Palaeogeography, Palaeoclimatology, Palaeoecology, v. 3, p. 29–48.
——— 1967c, Bering Land Bridge; evidence of spruce in late-Wisconsin times: Science, v. 156, p. 380–383.
Collier, A. J., 1902, A reconnaissance of the north-western portion of Seward Peninsula, Alaska: U.S. Geol. Survey Prof. Paper 2, 70 p.
Collins, F. R., 1958, Vegetated sand dunes in central Alaska: Geol. Soc. America Bull., v. 69, p. 1752.
Collins, H. B., Jr., 1937, Culture migrations and contacts in the Bering Sea region: Am. Anthropologist, v. 39, p. 375–384.
——— 1967, Introduction, *in* Giddings, J. L., Jr., Ancient men of the Arctic: New York, A. A. Knopf, p. 19–31.
Cooling, L. F., and Ward, W. H., 1944, Damage to cold stores due to frost heaving: Inst. Refrig. Proc., London, v. 41, p. 37–47.
Coonrad, W. L., 1957, Geologic reconnaissance in the Yukon-Kuskokwim delta region, Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I-223, scale 1:500,000.
Cooper, W. S., 1923, The recent ecological history of Glacier Bay, Alaska: Ecology, v. 4, p. 93–128, 223–246, 355–365.
——— 1931, The third expedition to Glacier Bay, Alaska: Ecology, v. 12, p. 61–95.
——— 1937, The problem of Glacier Bay, Alaska—a study of glacier variations: Geog. Rev., v. 27, p. 37–62.
——— 1939, A fourth expedition to Glacier Bay, Alaska: Ecology, v. 20, p. 130–155.
——— 1942, Vegetation of the Prince William Sound region, Alaska with a brief excursion into post-Pleistocene climatic history: Ecol. Monograph, v. 12, p. 1–22.
Coulter, H. W., and Coulter, E. B., 1961, Geology of the Valdez (A-5) quadrangle, Alaska: U.S. Geol. Survey Geol. Quad. Map GQ–142, scale 1:63,360.
——— 1962, Preliminary geologic map of the Valdez-Tiekel Belt, Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–356, scale 1:96,000.
Coulter, H. W., Hopkins, D. M., Karlstrom, T. N. V., Péwé, T. L., Wahrhaftig, Clyde, and Williams, J. R., 1965, Extent of glaciations in Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–415, scale 1:2,500,000.

Coulter, H. W., Hussey, K. M., and O'Sullivan, J. B., 1960, Radiocarbon dates relating to the Gubik formation, northern Alaska, *in* Short papers in the geological sciences 1960: U.S. Geol. Survey Prof. Paper 400–B, p. B350–B351.

Cox. Allan, 1969, Geomagnetic reversals: Science, v. 163, p. 237–245.

Cox, Allan, and Dalrymple, G. B., 1967, Geomagnetic polarity epochs; Nunivak Island, Alaska: Earth and Planetary Sci. Letters, v. 3, p. 173–177.

Cox, Allan, Doell, R. R., and Dalrymple, G. B., 1963a, Geomagnetic polarity epochs and Pleistocene geochronometry: Nature, v. 198, no. 4885, p. 1049–1051.

——— 1963b, Geomagnetic polarity epochs—Sierra Nevada II: Science, v. 142, no. 3590, p. 382–385.

——— 1964a, Geomagnetic polarity epochs: Science, v. 143, p. 351–352.

——— 1964b, Reversals of the earth's magnetic field: Science, v. 144, p. 1537–1543.

——— 1965, Quaternary paleomagnetic stratigraphy, *in* Wright, H. E., Jr., and Frey, D. G., eds., The Quaternary of the United States: Princeton, Princeton Univ. Press, p. 817–830.

Cox, Allan, Hopkins, D. M., and Dalrymple, C. B., 1966, Geomagnetic polarity epochs; Pribilof Islands, Alaska: Geol. Soc. America Bull., v. 77, p. 883–910.

Crane, H. R., 1956, University of Michigan radiocarbon dates I: Science, v. 124, p. 664–672.

Creager, J. S., and McManus, D. A., 1967, Geology of the floor of Bering and Chukchi Seas—American studies, *in* Hopkins, D. M., ed., The Bering Land Bridge: Stanford, Stanford Univ. Press, p. 7–31.

Curray, J. R., 1965, Late Quaternary history, continental shelves of the United States, *in* Wright, H. E., and Frey, D. G., eds., The Quaternary of the United States: Princeton, Princeton Univ. Press, p. 723–735.

Curtis, G. H., 1969, The stratigraphy of the ejecta from the 1912 eruption of Mount Katmai and Novarupta, Alaska, *in* Coats, R. R., Hay, R. L., and Anderson, C. A., eds., Studies in volcanology: Geol. Soc. America Mem. 116, p. 153–210.

Czudek, Tadeas, and Demek, Jaromir, 1970, Thermokarst and its influence on the development of lowland relief: Quaternary Research, v. 1, p. 103–120.

Dachnowski-Stokes, A. P., 1941, Peat resources in Alaska: U.S. Dept. Agriculture Tech. Bull. 769, 84 p.

Dall, W. H., 1869, On the bones of the musk ox and buffalo found in the Yukon: Boston Soc. Nat. History Proc., v. 13, p. 136–137.

——— 1873, Remarks on tusks of Elephas from Kotzebue Sound: California Acad. Sci. Proc., v. 4, pt. 5, p. 299–302.

——— 1884, Glaciation in Alaska: Philos. Soc. Washington, Bull. 6, p. 33–36.

——— 1885, Further notes on Bogoslof Island: Science, v. 5, p. 32–33.

——— 1896, Report on the coal and lignite of Alaska: U.S. Geol. Survey 17th Ann. Rept., pt. 1, p. 763–875.

——— 1905, Land fresh water mollusks (Alaska): Harriman Alaska Expedition, v. 13: New York, Doubleday, Page and Company, 191 p.

——— 1920, Pliocene and Pleistocene fossils from the Arctic Coast of Alaska and the auriferous beaches of Nome, Norton Sound, Alaska: U.S. Geol. Survey Prof. Paper 125, p. 23–37.

Dall, W. H., and Harris, G. D., 1892, Correlation papers, Neocene: U.S. Geol. Survey Bull. 84, 349 p.

Dansgaard, W., Johnsen, S. J., and Møller, J., 1969, One thousand centuries of climatic record from Camp Century on the Greenland ice sheet: Science, v. 166, no. 3903, p. 377–381.

Davidson, D. T., and Handy, R. L., 1952, Property variations in the Peorian loess of southwestern Iowa: Iowa Acad. Sci. Proc., v. 59, p. 248–265.

Davidson, D. T., Roy, C. J., and others, 1959, The geology and engineering characteristics of some Alaskan soils: Iowa State Univ. Bull. 186, 149 p.

Dawson, G. M., 1889, Report on an exploration in the Yukon District, N. W. T. and adjacent northern portion of British Columbia, 1887: Canada Geol. Survey Ann. Rept., pt. 1, rept. B, 183 p.

Dement, J. A., 1962, The morphology and genesis of the subarctic brown forest soils of central Alaska: Cornell Univ., Ph.D. thesis, 147 p.

Denton, G. H., and Armstrong, R. L., 1969, Miocene-Pliocene glaciations in southern Alaska: Am. Jour. Sci., v. 267, p. 1121–1142.

Denton, G. H., Armstrong, R. L., and Stuiver, Minze, 1969, Histoire glaciaire et chronologie de la région du détroit de McMurdo, sud de la Terre Victoria, Antarctide; note Préliminaire: Rev. Géographie Phys. et Géologie Dynam., v. 11, p. 265–278.

——— 1971, The late Cenozoic glacial history of Antarctica, *in* Turekian, K. K., ed., The late Cenozoic glacial ages: New Haven, Yale Univ. Press, p. 267–306.

Detterman, R. L., 1953, Sagavanirktok-Anaktuvuk region, northern Alaska, *in* Péwé, T. L., and others, Multiple glaciation in Alaska: U.S. Geol. Survey Circ. 289, p. 11–12.

——— 1970, Early Holocene warm interval in northern Alaska: Arctic, v. 23, p. 130–132.

Detterman, R. L., Bickel, R. S., and Gryc, George, 1963, Geology of the Chandler River region, Alaska: U.S. Geol. Prof. Paper 303–E, p. 223–324.

Detterman, R. L., Bowsher, A. L., and Dutro, J. T., Jr., 1958, Glaciation on the Arctic Slope of the Brooks Range, northern Alaska: Arctic, v. 11, p. 43–61.

Detterman, R. L., and Reed, B. L., 1967, Surficial deposits of the Iliamna quadrangle, Alaska: U.S. Geol. Survey open-file map, 1 sheet.

Donn, W. L., and Ewing, Maurice, 1966, A theory of ice ages, 3: Science, v. 152, p. 1706–1712.

Dorsh, J. B., 1934, Notes on the Pleistocene and Recent geology of the Fairbanks district, Alaska: Univ. Alaska, unpub. Bachelor's thesis, 31 p.

Douglas, L. A., and Tedrow, J. C. F., 1960, Tundra soils of arctic Alaska: Internat. Cong. Soil Sci., Madison 1960, Trans., v. 4, p. 291–304.

Drew, J. V., and Shanks, R. E., 1965, Landscape relationships of soils and vegetation in the forest-tundra ecotone, upper Firth River Valley, Alaska-Canada: Ecol. Monograph, v. 35, p. 285–306.

Dring, Nat, 1967, The Earth is your spaceship: Fort Worth, M. D. Evans Co., 113 p.

Drury, W. H., Jr., 1956, Bog flats and physiographic processes in the Upper Kuskokwim River region, Alaska: Cambridge, Gray Herbarium, Harvard Univ. Contr. 178, 130 p.

Dumond, D. E., 1965, On Eskaleutian linguistics, archaeology, and prehistory: Am. Anthropologist, v. 67, no. 5, p. 1231–1257.

Durham, J. W., and MacNeil, F. S., 1967, Cenozoic migrations of marine invertebrates through the Bering Strait region, *in* Hopkins, D. M., ed., The Bering Land Bridge: Stanford, Stanford Univ. Press, p. 326–349.

Eakin, H. M., 1916, The Yukon-Koyukuk region, Alaska: U.S. Geol. Survey Bull. 631, 88 p.

——— 1918, The Cosna-Nowitna region, Alaska: U.S. Geol. Survey Bull. 667, 54 p.

Eardley, A. J., 1938, Unconsolidated sediments and topographic features of the lower Yukon valley (Alaska): Geol. Soc. America Bull., v. 49, p. 303–341.

Einarsson, Thorleifur, Hopkins, D. M., and Doell, R. R., 1967, The stratigraphy of Tjornes, Northern Iceland, and the history of the Bering Land Bridge, *in* Hopkins, D. M., ed., The Bering Land Bridge: Stanford, Stanford Univ. Press, p. 312–325.

Everett, K. R., 1962, Quantitative analysis of slope movement in northwestern Alaska: Geol. Soc. America Spec. Paper 68, p. 115.
——— 1966, Instruments for measuring mass-wasting, *in* Internat. Permafrost Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 136–139.
Ewing, W. M., and Donn, W. L., 1956, A theory of ice ages: Science, v. 123, p. 1061–1066.
——— 1958, A theory of ice ages, 2: Science, v. 127, p. 1159–1162.
Faas, R. W., 1962, Foraminiferal paleoecology of the Gubik (Pleistocene) Formation of the Barrow Area, northern Alaska: Iowa Acad. Sci. Proc., v. 69, p. 354–361.
——— 1966, Paleoecology of an arctic estuary: Arctic, v. 19, p. 343–348.
Fairbridge, R. W., ed., 1968, The encyclopedia of geomorphology: New York, Reinhold Book Corp., 1240 p.
Farrand, W. R., 1961, Frozen mammoths and modern geology: Science, v. 133, p. 729–735.
Fenner, C. N., 1920, The Katmai Region, Alaska, and the great eruption of 1912: Jour. Geology, v. 28, p. 569–606.
——— 1923, The origin and mode of emplacement of the great tuff deposit in the Valley of Ten Thousand Smokes: Natl. Geog. Soc., Contributed Tech. Paper, Katmai Series, no. 1, 74 p.
Fernald, A. T., 1960, Geomorphology of the upper Kuskokwim region, Alaska: U.S. Geol. Survey Bull. 1071–G, p. 191–279.
——— 1962, Radiocarbon dates relating to a widespread volcanic ash deposit, eastern Alaska, *in* Short papers in geology, hydrology, and topography 1962: U.S. Geol. Survey Prof. Paper 450–B, p. B29–B30.
——— 1964, Surficial geology of the central Kobuk River valley, northwestern Alaska: U.S. Geol. Survey Bull. 1181–K, 31 p.
——— 1965a, Glaciation in the Nabesna River area, upper Tanana River valley, Alaska, *in* Geological Survey research 1965: U.S. Geol. Survey Prof. Paper 525–C, p. C120–C123.
——— 1965b, Recent history of the upper Tanana River lowland, Alaska, *in* Geological Survey research 1965: U.S. Geol. Survey Prof. Paper 525–C, p. C124–C127.
Fernald, A. T., and Nichols, D. R., 1953, Active sand dunes in the Kobuk River valley, northwestern Alaska [abs.]: Geol. Soc. America Bull., v. 64, p. 1421.
Ferrians, O. J., Jr., 1963a, Glaciolacustrine diamicton deposits in the Copper River Basin, Alaska, *in* Short papers in geology and hydrology 1963: U.S. Geol. Survey Prof. Paper 475–C, p. C121–C125.
——— 1963b, Till-like glaciolacustrine deposits in the Copper River Basin, Alaska [abs.]: Geol. Soc. America Spec. Paper 73, p. 151.
——— 1965a, Distribution and character of permafrost in the discontinuous permafrost zone of Alaska (summary), *in* Canadian Regional Permafrost Conf., 1964, Proc.: Natl. Research Council Canada Assoc. Comm. Soil and Snow Mechanics Tech. Memo 86, p. 15–16.
——— 1965b, Permafrost map of Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–445, scale 1:2,500,000.
——— 1966, Permafrost map of Alaska, *in* Internat. Permafrost Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 172.
Ferrians, O. J., Jr., Kachadoorian, Reuben, and Greene, G. W., 1969, Permafrost and related engineering problems in Alaska: U.S. Geol. Survey Prof. Paper 678, 37 p.
Ferrians, O. J., Jr., and Nichols, D. R., 1965, Copper River Basin, *in* Péwé, T. L., Ferrians, O. J., Jr., Nichols, D. R., and Karlstrom, T. N. V., Guidebook for Field Conf. F, Central and south central Alaska—Internat. Assoc. Quaternary Research, 7th Cong., USA 1965: Lincoln, Nebr., Nebraska Acad. Sci., p. 93–114.
Ferrians, O. J., Jr., Nichols, D. R., and Schmoll, H. R., 1958, Pleistocene volcanic mudflow in the Copper River basin, Alaska [abs.]: Geol. Soc. America Bull., v. 69, p. 1563.
Ferrians, O. J., Jr., and Schmoll, H. R., 1957, Extensive proglacial lake of Wisconsin age in the Copper River basin, Alaska [abs.]: Geol. Soc. America Bull., v. 68, p. 1726.
Feulner, A. J., and Schupp, R. G., 1964, Temperature and chemical quality of water from a well drilled through permafrost near Bethel, Alaska: U.S. Geol. Survey Prof. Paper 501–D, p. D144–D148.
Field, W. O., Jr., 1932a, The mountains and glaciers of Prince William Sound: Am. Alpine Jour., v. 1, p. 445–458.
——— 1932b, The glaciers of the northern part of Prince William Sound, Alaska: Geog. Rev., v. 22, p. 361–368.
——— 1941, A gazeteer of Alaska glaciers: Am. Geophys. Union Trans., 22d Ann. Mtg., pt. 3, p. 796–799.
——— 1942, Glacier studies in Alaska, 1941: Geog. Rev., v. 32, p. 154–155.
——— 1947, Glacier recession in Muir Inlet, Glacier Bay, Alaska: Geog. Rev., v. 37, p. 369–399.
——— 1950, The variations of Alaskan glaciers, 1935–1947: Internat. Assoc. Sci. Hydrology, Assemblie Générale, Oslo, v. 2, Comm. de la Neige et des Glaciers, Gravaux, p. 277–282.
——— 1965, Avalanches caused by the Alaska earthquake of March 1964, *in* International symposium on scientific aspects of snow and ice avalanches: Internat. Assoc. Sci. Hydrology Pub. 69, p. 326–331.
——— 1968, The effect of previous earthquakes on glaciers, *in* The great Alaska earthquake of 1964: Natl. Acad. Sci. Pub. 1603, p. 252–265.
Field, W. O., Jr., and Miller, M. M., 1950, The Juneau Ice Field research project: Geog. Rev., v. 40, p. 179–190.
——— 1951, Studies of the Taku Glacier, Alaska: Jour. Geology, v. 59, p. 622–623.
Field, W. O., Jr., Sater, J. E., and Ragle, R. H., 1964, Effects on glaciers of the Alaskan earthquakes of March 1964; report of an aerial reconnaissance, 14 to 19 April 1964: Arctic Inst. North America Research Paper 28, 23 p.
Fisk, H. N., and McFarlan, E., Jr., 1955, Late Quaternary deltaic deposits of the Mississippi River: Geol. Soc. America Spec. Paper 62, p. 279–302.
Flerow, C. C., 1967, On the origin of the mammalian fauna of Canada, *in* Hopkins, D. M., ed., The Bering Land Bridge: Stanford, Stanford Univ. Press, p. 271–280.
Flint, R. F., 1957, Glacial and Pleistocene geology: New York, John Wiley & Sons, 553 p.
——— 1965, The Pliocene-Pleistocene boundary: Geol. Soc. America Spec. Paper 84, p. 497–533.
——— 1971, Glacial and Quaternary geology: New York, John Wiley & Sons, 892 p.
Foster, H. L., 1967, Geology of the Mount Fairplay area, Alaska: U.S. Geol. Survey Bull. 1241–B, 18 p.
Foster, H. L., and Holmes, G. W., 1965, A large transitional rock glacier in the Johnson River area, Alaska Range, *in* Geological Survey research 1965: U.S. Geol. Survey Prof. Paper 525–B, p. B112–B116.
Frick, Childs, 1930, Alaska's frozen fauna: Nat. History, v. 30, p. 71–80.
——— 1933, New remains of trilophodont-terabelodont mastodons: Am. Mus. Nat. History Bull., v. 59, p. 505–652.
——— 1937, Horned ruminants of North America: Am. Mus. Nat. History Bull., v. 69, 669 p.
Fyles, J. G., Henginbottom, J. A., and Rampton, V. N., 1972, Quater-

nary geology and geomorphology, MacKenzie Delta to Hudson Bay: Internat. Geol. Cong. Guidebook 24th Sess., Montreal, Field Excursion A30, 23 p.
Gard, L. M., Jr., Lewis, G. E., and Whitmore, F. C., Jr., 1972, Steller's Sea Cow in Pleistocene interglacial beach deposits on Amchitka, Aleutian Islands: Geol. Soc. America Bull., v. 83, p. 867–870.
Gard, L. M., and Szabo, B. J., 1971, Age of the Pleistocene deposits at South Bight, Amchitka Island, Alaska [abs.]: Geol. Soc. America, Ann. Mtg., Abs. with Programs, v. 3, no. 7, p. 577.
Geist, O. W., 1940, Mammoth's foot found: Alaska Univ., Farthest-North Collegian, 1 p.
——— 1953, Collecting Pleistocene fossils in Alaska: Alaskan Sci. Conf., 2d, Mt. McKinley Natl. Park 1951, Proc., p. 171–172.
Geist, O. W., and Péwé, T. L., 1957, Quantitative measurements of the 1937 advance of the Black Rapids Glacier, Alaska [abs.]: Alaskan Sci. Conf., 5th, Anchorage 1954, Proc., p. 51–52.
Geist, O. W., and Rainey, F. G., 1936, Archaeological excavations at Kukulik, St. Lawrence Island, Alaska: Alaska Univ., Misc. Pub., v. 2, 391 p.
Gerard, R. D., 1954, Some aspects of Pleistocene and post-glacial climate change in central Alaska: McGill Univ., M. S. thesis, 24 p.
Giddings, J. L., Jr., 1938, Buried wood from Fairbanks, Alaska: Tree-ring Bull., v. 4, p. 3–6.
——— 1951, The Denbigh Flint complex: Am. Antiquity, v. 16, p. 193–203.
——— 1960, The archeology of Bering Strait: Current Anthropology, v. 1, no. 2, p. 121–138.
——— 1962, Seven discoveries of Bering Strait: Am. Philos. Soc., v. 106, p. 89–93.
——— 1964, The archaeology of Cape Denbigh: Providence, Brown Univ. Press, 331 p.
——— 1967, Ancient men of the Arctic: New York, A. A. Knopf, 391 p.
Gilbert, G. K., 1910, Glaciers and glaciation: Harriman Alaska Expedition: New York, Doubleday, Page and Company, 231 p.
Gillespie, J. M., 1970, A comparison of the proteins in the hairs of the mammoth and the elephant: Science, v. 170, 1100–1102.
Gilmore, C. W., 1908, Smithsonian exploration in Alaska in 1907 in search of Pleistocene fossil vertebrates: Smithsonian Misc. Colln., v. 51, p. 1–38.
Goldthwait, R. P., 1963, Dating the Little Ice Age in Glacier Bay, Alaska: Internat. Geol. Cong., 21st, Norden 1960, Proc., pt. 27, p. 37–46.
——— 1966, Evidence from Alaskan glaciers of major climatic changes, *in* World climate from 8000 to 0 B.C., Internat. symposium on world climate, London 1966, Proc.: London, Royal Meteorol. Soc., p. 40–53.
Grant, U. S., and Higgins, D. F., 1913, Coastal glaciers of Prince William Sound and Kenai Peninsula, Alaska: U.S. Geol. Survey Bull. 526, 75 p.
Greene, G. W., Lachenbruch, A. H., and Brewer, M. C., 1960, Some thermal effects of a roadway on permafrost, *in* Short papers in the geological sciences 1960: U.S. Geol. Survey Prof. Paper 400–B, p. B141–B144.
Griggs, R. F., 1922, The Valley of Ten Thousand Smokes (Alaska): Washington, Natl. Geog. Soc., 340 p.
——— 1936, The vegetation of the Katmai district: Ecology, v. 17, p. 380–417.
Grim, M. S., and McManus, D. S., 1970, A shallow seismic-profiling survey of the northern Bering Sea: Marine Geology, v. 8, p. 293–320.
Gryc, George, Patton, W. W., Jr., and Payne, T. G., 1951, Present Cretaceous stratigraphic nomenclature of northern Alaska: Washington Acad. Sci. Jour., v. 41, p. 159–167.
Guthrie, R. D., 1966a, Bison horn cores—character choice and systematics: Jour. Paleontology, v. 40, no. 3, p. 738–740.
——— 1966b, The extinct Wapiti of Alaska and Yukon Territory: Canada Jour. Zoology, v. 44, p. 47–57.
——— 1966c, Pelage of fossil bison—a new osteological index: Jour. Mammalogy, v. 47, no. 4, p. 725–727.
——— 1967, Differential preservation and recovery of Pleistocene large mammal remains in Alaska: Jour. Paleontology, v. 41, no. 1, p. 243–246.
——— 1968a, Paleoecology of the large-mammal community in interior Alaska during the Late Pleistocene: Am. Midland Naturalist, v. 79, no. 2, p. 346–363.
——— 1968b, Paleoecology of a late Pleistocene small mammal community from interior Alaska: Arctic, v. 21, no. 4, p. 223–244.
——— 1972, Re-creating a vanished world: Natl. Geog. Mag., v. 141, no. 3, p. 295–301.
Guthrie, R. D., and Matthews, J. V., Jr., 1971, The Cape Deceit fauna—early Pleistocene mammalian assemblage from the Alaskan Arctic: Quaternary Research, v. 1, no. 4, p. 474–510.
Hackman, R. J., 1965, Interpretation of Alaskan post-earthquake photographs: Photogramm. Eng., v. 31, p. 604–610.
Hadleigh-West, Frederick, 1963, ed., Early man in the western American Arctic: Alaska Univ., Anthropol. Papers, v. 10, 135 p.
——— The Donnelly Ridge site and the definition of an early core and blade complex in central Alaska: Am. Antiquity, v. 32, no. 3, p. 360–382.
Hamelin, L. E., and Cook, F. A., 1967, Illustrated glossary of periglacial phenomenon: Quebec, Univ. Laval Press, 237 p.
Hamilton, T. D., 1965a, Alaskan temperature fluctuations and trends; analysis of recorded data: Arctic, v. 18, p. 105–117.
——— 1965b, Comparative glacier photographs from northern Alaska: Jour. Glaciology, v. 5, no. 40, p. 479–487.
——— 1966, Geomorphology and glacial history of the Alatna Valley, northern Alaska: Washington Univ., Ph.D. dissertation, 264 p.
——— 1969, Glacial geology of the lower Alatna Valley, Brooks Range, Alaska: Geol. Soc. America Spec. Paper 123, p. 181–223.
Hance, J. H., 1937, The recent advance of Black Rapids Glacier, Alaska: Jour. Geology, v. 45, p. 775–783.
Hansen, H. P., 1953, Postglacial forests in the Yukon Territory and Alaska: Am. Jour. Sci., v. 251, p. 505–542.
Hansen, W. R., 1965, Effects of the earthquake of March 27, 1964, at Anchorage, Alaska: U.S. Geol. Survey Prof. Paper 542–A, 68 p.
Hanson, L. W., 1965, Size distribution of the White River Ash, Yukon Territory: Alberta Univ., M.S. thesis, 59 p.
Hapgood, C. H., 1958, Earth's shifting crust: New York, Pantheon Books, 483 p.
——— 1960, The mystery of the frozen mammoths: Coronet (September), p. 76.
Hardy, R. M., and D'Appolonia, Elio, 1946, Permanently frozen ground and foundation design: Eng. Jour. Canada, v. 29, p. 1–11.
Harington, C. R., 1969, Pleistocene remains of the lion-like cat (*Panthera atrox*) from the Yukon Territory and northern Alaska: Canadian Jour. Earth Sci., v. 6, p. 1277–1288.
——— 1970a, Ice age mammal research in the Yukon Territory and Alaska, *in* Early man and environments in northwest North America: Calgary Univ. Archeol. Assoc., p. 35–51.
——— 1970b, A Pleistocene muskox (*Ovibos moschatus*) from gravels of Illinoian age near Nome, Alaska: Canadian Jour. Earth Sci., v. 7, no. 5, p. 1326–1331.
Harrington, G. L., 1918, The Anvik-Andreafski region, Alaska (including the Marshall district): U.S. Geol. Survey Bull. 683, 70 p.
Haselton, G. M., 1966, Glacial geology of Muir Inlet, southeast Alaska: Ohio State Univ. Inst. Arctic Studies Rept. 18, 34 p.

Hastenrath, S. L., 1967, Observations on the snowline in the Peruvian Andes: Jour. Glaciology, v. 6, no. 46, p. 541–550.

——— 1971, On the Pleistocene snowline depression in the arid regions of the South American Andes: Jour. Glaciology, v. 10, no. 59, p. 255–267.

Hayes, C. W., 1892, An expedition through the Yukon district: Natl. Geog. Mag., v. 4, p. 117–162.

Herman, Yvonne, 1970, Arctic paleo-oceanography in late Cenozoic time: Science, v. 169, p. 474–477.

Herman, Yvonne, Grazzini, C. V., and Hooper, C., 1971, Arctic palaeotemperatures in late Cenozoic time: Nature, v. 232, p. 466–469.

Hershey, O. H., 1909, The ancient Kobuk glacier of Alaska: Jour. Geology, v. 17, p. 83–91.

Herz, O. F., 1904, Frozen mammoth in Siberia: Smithsonian Ann. Rept., 1903, p. 611–625.

Heusser, C. J., 1952, Pollen profiles from southeastern Alaska: Ecol. Monograph, v. 22, p. 331–352.

——— 1958, Late-Pleistocene environments and chronology of Pacific coastal Alaska [abs.]: Geol. Soc. America Bull., v. 69, p. 1753–1754.

——— 1960, Late-Pleistocene environments of North Pacific North America: Am. Geog. Soc. Spec. Pub. 35, 308 p.

——— 1963a, Postglacial palynology and archaeology in the Naknek River drainage area, Alaska: Am. Antiquity, v. 29, p. 74–81.

——— 1963b, Pollen diagrams from Ogotoruk Creek, Cape Thompson, Alaska: Grana Palynologica, v. 4, p. 149–159.

——— 1966, Pleistocene climatic variations in the western United States, *in* Blumenstock, D. I., ed., Pleistocene and post-Pleistocene climatic variations in the Pacific area—a symposium: Honolulu, Bishop Mus. Press, p. 9–36.

Heusser, C. J., and Marcus, M. G., 1964, Historical variations of Lemon Creek Glacier, Alaska, and their relationships to the climatic record: Jour. Glaciology, v. 5, p. 77–86.

Hibbard, C. W., Ray, C. E., Savage, D. E., Taylor, D. W., and Guilday, J. E., 1965, Quaternary mammals of North America, *in* Wright, H. E., Jr., and Frey, D. G., eds., The Quaternary of the United States: Princeton, Princeton Univ. Press, p. 509–525.

Hibben, F. C., 1943, Evidence of early man in Alaska: Am. Antiquity, v. 8, p. 254–259.

Higashi, Akira, 1958, Experimental study of frost heaving: U.S. Army Corps Engineers, Snow, Ice and Permafrost Research Establishment, Research Rept. 45, 20 p.

Hill, J. E., 1942, *Citellus parryi* from the Pleistocene of Alaska [abs.]: Geol. Soc. America Bull., v. 53, p. 1842.

Hoare, J. M., and Condon, W. H., 1971, Geology map of the Marshall quadrangle, western Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–668, scale 1:250,000.

Hoare, J. M., Condon, W. H., Cox, Allan, and Dalrymple, G. B., 1968, Geology, paleomagnetism, and potassium-argon ages of basalts from Nunivak Island, Alaska: Geol. Soc. America Mem. 116, p. 377–413.

Hoare, J. M., and Coonrad, W. L., 1959, Geology of the Russian Mission quadrangle, Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–292, scale 1:250,000.

———1961a, Geologic map of the Goodnews quadrangle, Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–339, scale 1:250,000.

——— 1961b, Geologic map of the Hagemeister Island quadrangle, Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–321, scale 1:250,000.

Holmes, G. W., 1959a, Geological and hydrological investigations at Lake Peters, *in* Hartshorn, J. H., Proceedings of the First Annual Arctic Planning Session, Nov. 1958: U.S. Air Force Cambridge Research Center, Geophysics Research Directorate Research Note 15, p. 2–6.

——— 1959b, Glacial geology of the Mt. Michelson B–2 quadrangle, Alaska, *in* Preliminary report of the Mt. Chamberlain–Barter Island Project, Alaska: U.S. Geol. Survey open-file report, p. 47–60.

——— 1959c, Glaciation in the Johnson River–Tok area, Alaska: Geol. Soc. America Bull., v. 70, p. 1620.

——— 1965, Geologic reconnaissance along the Alaska Highway, Delta River to Tok Junction, Alaska: U.S. Geol. Survey Bull. 1181–H, 19 p.

Holmes, G. W., and Benninghoff, W. S., 1957, Terrain study of the Army test area, Fort Greeley, Alaska: U.S. Geol. Survey, Military Geology Br. (for U.S. Army Corps Engineers, Waterways Expt. Sta., Vicksburg, Miss.), 287 p.

Holmes, G. W., and Foster, H. L., 1968, Geology of the Johnson River area, Alaska: U.S. Geol. Survey Bull. 1249, 49 p.

Holmes, G. W., Foster, H. L., and Hopkins, D. M., 1966, Distribution and age of pingos of interior Alaska, *in* Internat. Permafrost Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 88–98.

Holmes, G. W., Hopkins, D. M., and Foster, H. L., 1968, Pingos in central Alaska: U.S. Geol. Survey Bull. 1241–H, 40 p.

Holmes, G. W., and Lewis, C. R., 1961, Glacial geology of the Mount Chamberlain area, Brooks Range, Alaska, *in* Raasch, G. O., ed., Geology of the Arctic: Toronto, Univ. Toronto Press, v. 2, p. 848–864.

——— 1965, Quaternary geology of the Mount Chamberlin area, Brooks Range, Alaska: U.S. Geol. Survey Bull. 1201–B, 32 p.

Holmes, G. W., and Péwé, T. L., 1965, Geologic map of the Mount Hayes D–3 quadrangle, Alaska: U.S. Geol. Survey Geol. Quad. Map GQ–366, scale 1:63,360.

Hooker, D. E., 1958, Those astounding ice ages: New York, Exposition Press, 148 p.

Hopkins, D. M., 1949, Thaw lakes and thaw sinks in the Imuruk Lake area, Seward Peninsula, Alaska: Jour. Geology, v. 57, p. 119–131.

——— 1959a, Some characteristics of the climate in forest and tundra regions in Alaska: Arctic, v. 12, p. 215–220.

——— 1959b, Cenozoic history of the Bering land bridge: Science, v. 129, p. 1519–1528.

——— 1963, Geology of the Imuruk Lake area, Seward Peninsula, Alaska: U.S. Geol. Survey Bull. 1141–C, 101 p. [1964].

——— 1965, Chetvertichnyye morskie transgresii Alyaski [Quaternary marine transgressions in Alaska], *in* Markov, F. G., and others, eds., Anthropogene period in Arctic and Subarctic, U.S.S.R.: Naucho-Issled. Inst. Geologii Arktiki Trudy, v. 143, p. 131–154 (English abstract).

——— 1967a, The Bering Land Bridge: Stanford, Stanford Univ. Press, 495 p.

——— 1967b, Quaternary marine transgressions in Alaska, *in* Hopkins, D. M., ed., The Bering Land Bridge: Stanford, Stanford Univ. Press, p. 47–90.

——— 1970, Paleoclimatic speculations suggested by new data of the location of the spruce refugium in Alaska during the last glaciation, *in* Abstracts: Am. Quaternary Assoc., 1st Mtg., p. 67.

——— 1972, The paleogeography and climatic history of Beringia during late Cenozoic time: Internord, no. 12, p. 121–150.

Hopkins, D. M., and Einarsson, Thorleifur, 1966, Pleistocene glaciation on St. George, Pribilof Islands: Science, v. 152, p. 343–345.

Hopkins, D. M., and Giddings, J. L., Jr., 1953, Geological background of the Iyatayet archaeological site, Cape Denbigh, Alaska: Smithsonian Misc. Colln., v. 121, no. 11, 33 p.

Hopkins, D. M., Karlstrom, T. N. V., and others, 1955, Permafrost and ground water in Alaska: U.S. Geol. Survey Prof. Paper 264–F, p.

113–146.
Hopkins, D. M., McCulloch, D. S., and Janda, R. J., 1962, Pleistocene structure of Baldwin Peninsula, Kotzebue Sound, Alaska [abs.]: Geol. Soc. America Spec. Paper 68, p. 116.
Hopkins, D. M., and MacNeil, F. S., 1960, A marine fauna probably of late Pliocene age near Kivalina, Alaska, *in* Short papers in the geological sciences 1960: U.S. Geol. Survey Prof. Paper 400–B, p. B339–B342.
Hopkins, D. M., MacNeil, F. S., and Leopold, E. B., 1960, The coastal plain at Nome, Alaska; a late-Cenozoic type section for the Bering Strait region: Internat. Geol. Cong., 21st, Copenhagen 1960, Proc., pt. 4, p. 46–57.
Hopkins, D. M., MacNeil, F. S., Merklin, R. L., and Petrov, O. M., 1965, Quaternary correlations across Bering Strait: Science, v. 147, p. 1107–1114.
Hopkins, D. M., Matthews, J. V., Wolfe, J. A., and Silberman, M. L., 1971, A Pliocene flora and insect fauna from the Bering Strait region: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 9, p. 211–231.
Hopkins, D.M., Rowland, R. W., Echols, R. E., and Valentine, P. C., 1974, Anvilian (early Pleistocene) marine fauna from western Seward Peninsula, Alaska: Quaternary Research, v. 4, p. 441–470.
Hopkins, D. M., Rowland, R. W., and Patton, W. W., Jr., 1972, Middle Pleistocene mollusks from St. Lawrence Island and their significance for the paleo-oceanography of the Bering Sea: Quaternary Research, v. 2, no. 2, p. 119–134.
Hopkins, D. M., Scholl, D. W., Addicott, W. O., Pierce, R. L., Smith, P. B., Wolfe, J. A., Gershanovich, D., Kotenev, B., Lohman, K. E., Lipps, J. H., and Obradovich, J., 1969, Cretaceous Tertiary and early Pleistocene rocks from the continental margin in the Bering Sea: Geol. Soc. America Bull., v. 80, no. 8, p. 1471–1480.
Hopkins, D. M., and Sigafoos, R. S., 1951, Frost action and vegetation patterns on Seward Peninsula, Alaska: U.S. Geol. Survey Bull. 974–C, p. 51–101.
——— 1954, Role of frost thrusting in the formation of tussocks: Am. Jour. Sci., v. 252, p. 55–59.
Hoskin, C. M., Guthrie, R. D., and Hoffman, B. L. P., 1970, Pleistocene and Recent bird gastroliths from interior Alaska: Arctic, v. 23, no. 1, p. 14–23.
Hosley, Edward, 1969, *Comments on* Bryan, A. L., 1969, Early man in America in the late-Pleistocene chronology of western Canada and Alaska: Current Anthropology, v. 10, p. 355.
Howitt, Frank and Clegg, M. W., 1970, Permafrost in the Prudhoe Bay field; geology and physical characteristics [abs.]: Am. Assoc. Petroleum Geologists Bull., v. 54, no. 12, p. 2487.
Howorth, H. H., 1887, The mammoth and the flood: London, S. Low, Marston, Searle, and Rivington, Ltd., 464 p.
Hubley, R. C., 1955, Measurements of diurnal variations in snow albedo on Lemon Creek Glacier, Alaska: Jour. Glaciology, v. 2, p. 560–563.
——— 1957, Analysis of surface energy during the ablation season on Lemon Creek Glacier: Am. Geophys. Union Trans., v. 38, p. 68–85.
Hughes, O. L., 1969, Distribution of open-system pingos in central Yukon territory with the respect to glacial limit: Canada Geol. Survey Paper 69–34, 8 p.
Hughes, O. L., Rampton, V. N., and Rutter, N. W., 1972, Quaternary geology and geomorphology, southern and central Yukon (northern Canada): Internat. Geol. Cong. Guidebook 24th Sess., Montreal, Field Excursion A 11, 59 p.
Hultén, Eric, 1937, Outline of the history of Arctic and boreal biota during the Quaternary period: Stockholm, Bokförlags aktiebolaget Thule, 168 p.
——— 1941–1948, Flora of Alaska and Yukon, part 1: Lunds Univ. Arssk, N. F. avd. 2, Band 37, no. 1, p. 1–127; part 2, idem, Band 38, no. 1, p. 132–412; part 3, idem, Band 39, no. 1, p. 415–567; part 4, idem, Band 40, no. 1, p. 571–795; part 5, idem, Band 41, no. 1, p. 799–978; part 6, idem, Band 42, no. 1, p. 981–1066; part 7, idem, Band 43, no. 1, p. 1069–1200; part 8, idem, Band 44, no. 1, p. 1203–1341.
Hunkins, K., Be., A. W. H., Opdyke, N. D., and Mathieu, G., 1971, The late Cenozoic history of the Arctic Ocean, *in* Turekian, K. K., ed., The late Cenozoic glacial ages: New Haven, Yale Univ. Press, p. 215–237.
Hussey, K. M., and Michelson, R. W., 1966, Tundra relief features near Point Barrow, Alaska: Arctic, v. 19, p. 162–184.
Ives, J. D., 1960, Semi-permanent snow beds in central Labrador-Ungava: McGill Univ. Sub-Arctic Research Paper No. 9, p. 45–56.
Ives, P. C., Levin, Betsy, Robinson, R. D., and Meyer, Rubin, 1964, U.S. Geological Survey radiocarbon dates 7: Radiocarbon v. 6, p. 37–76.
Jackson, K. A., and Chalmers, Bruce, 1956, Study of ice formation in soils: U.S. Army Corps Engineers Tech. Rept. 65, 29 p.
Jahn, Alfred, 1966, Alaska: Warsaw, Pánstwowe Wydawnictwo Naukowe, 498 p.
James, Nelson, and Sutherland, M. L., 1942, Are there living bacteria in permanently frozen subsoil?: Canadian Jour. Research, sec. C, v. 20, no. 6, p. 228–235.
Jenness, J. L., 1949, Permafrost in Canada: Arctic, v. 2, p. 13–27.
Johnson, A. W., 1952, Frost action in roads and airfields—a review of the literature 1765–1951: Highway Research Board Spec. Rept. 1, 287 p.
Johnson, Frederick, 1946, An archaeological survey along the Alaska Highway, 1944: Am. Antiquity, v. 11, p. 183–186.
Johnston, W. A., 1933, Quaternary geology of North America in relation to the migration of man, *in* Jenness, Diamond, ed., 1933, The American Aborigines, their origin and antiquity: Toronto, Univ. Toronto Press, p. 9–46.
Journaux, M. A., 1969, Phénomènes périglaciaires dans la Nord de l'Alaska et du Yukon: Géog. Assoc. France Bull. 368–369, p. 337–350.
Judson, Sheldon, 1946, Late-glacial and postglacial chronology on Adak: Jour. Geology, v. 54, p. 376–385.
Juhle, R. W., and Coulter, H. W., 1955, The Mount Spurr eruption, July 9, 1953: Am. Geophys. Union Trans., v. 36, p. 199–202.
Jumikis, A. R., 1955, The frost penetration problem in highway engineering: New Jersey, Rutgers Univ. Press, 162 p.
Kachadoorian, Reuben, Hopkins, D. M., and Nichols, D. R., 1954, A preliminary report of geologic factors affecting highway construction in the area between the Susitna and MacLaren Rivers, Alaska: U.S. Geol. Survey open-file report, 74 p.
Kahlke, H. D., 1963, Ovibos aus den Kiesen von Süssenborn. Ein Beitrag zur Systematic und Phylogenie der Ovibovini und zur Stratigraphie des Pleistozäns: Geologie, v. 12, no. 8, Berlin.
Karlstrom, T. N. V., 1955, Kenai Lowland, *in* Hopkins, D. M., Karlstrom, T.N.V., and others, Permafrost and ground water in Alaska: U.S. Geol. Survey Prof. Paper 264–F, p. 133–134.
——— 1957, Tentative correlation of Alaskan glacial sequences: Science, v. 125, p. 73–74.
——— 1960a, The Cook Inlet, Alaska, glacial record and Quaternary classification, *in* Short papers in the geological sciences 1960: U.S. Geol. Survey Prof. Paper 400–B, p. B330–B332.
——— 1960b, Surficial deposits of Alaska, *in* Short papers in the geological sciences 1960: U.S. Geol. Survey Prof. Paper 400–B, p. B333–B335.
——— 1961, The glacial history of Alaska; its bearing on paleoclimatic theory: New York Acad. Sci. Annals, v. 95, p. 290–340.

——— 1964, Quaternary geology of the Kenai lowland and glacial history of the Cook Inlet region, Alaska: U.S. Geol. Survey Prof. Paper 443, 69 p.

——— 1965, Upper Cook Inlet area and Matanuska River valley, *in* Péwé, T. L., Ferrians, O. J., Jr., Nichols, D. R., and Karlstrom, T. N. V., Guidebook for Field Conf. F, Central and south central Alaska—Internat. Assoc. Quaternary Research, 7th Cong., USA 1965: Lincoln, Nebr., Nebraska Acad. Sci., p. 114–141.

——— 1968, The Quaternary time scale—a current problem of correlation and radiometric dating, *in* Morrison, R. B., and Wright, H. E., Jr., eds., Means of correlation of Quaternary successions: Salt Lake City, Univ. Utah Press, p. 121–150.

——— 1969, Regional setting and geology, *in* Karlstrom, T. N. V., and Ball, G. E., eds., The Kodiak Island refugium; its geology, flora, fauna, and history: Edmonton, Alberta Univ., Boreal Inst., p. 20–54.

Karlstrom, T. N. V., and others, 1964, Surficial geology of Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–357, scale 1:1,584,000.

Kaye, C. A., 1961, Pleistocene stratigraphy of Boston, Massachusetts, *in* Short papers in the geologic and hydrologic sciences 1961: U.S. Geol. Survey Prof. Paper 424–B, p. B73–B75.

Kaye, C. A., and Barghoorn, E. S., 1964, Late Quaternary sea-level change and crustal rise at Boston, Massachusetts, the notes on the autocompaction of peat: Geol. Soc. America Bull., v. 73, p. 63–80.

Keeler, C. M., 1958, Ablation studies; lower McCall Glacier, June 23 to Sept. 1, 1957, *in* IGY Glaciological Rept., ser. 1: Am. Geog. Soc. 12, p. 1–15.

——— 1959, Notes on the geology of the McCall Valley area: Arctic, v. 12, p. 87–97.

Kent, Dennis, Opdyke, N. D., and Ewing, Maurice, 1971, Climate change in the North Pacific using ice-rafted detritus as a climatic indicator: Geol. Soc. America Bull., v. 82, p. 2741–2754.

Kind, N. V., 1967, Radiocarbon chronology in Siberia, *in* Hopkins, D. M., ed., The Bering Land Bridge: Stanford, Stanford Univ. Press, p. 172–192.

King, P. B., 1965, Tectonics of Quaternary time in middle North America, *in* Wright, H. E., Jr., and Frey, D. G., eds., The Quaternary of the United States: Princeton, Princeton Univ. Press, p. 831–232.

Kjøller, Annelise, and Ødum, Soren, 1971, Evidence for longevity of seeds and microorganisms in permafrost: Arctic, v. 24, no. 3, p. 230–232.

Klein, D. R., 1965, Postglacial distribution patterns of mammals in the southern coastal regions of Alaska: Arctic, v. 18, no. 1, p. 7–19.

Kotzebue, Otto von, 1821, A voyage of discovery into the South Sea and Behring's Straits, for the purpose of exploring a northeast passage, undertaken in the years 1815–1818: London, Longman, Hurst, Rees, Orme, and Brown, v. 1, 358 p.

Krinitsky, E. L., and Turnbull, W. J., 1967, Loess deposits of Mississippi: Geol. Soc. America Spec. Paper 94, 64 p.

Krinsley, D. B., 1965, Birch Creek pingo, Alaska, *in* Geological Survey research 1965: U.S. Geol. Survey Prof. Paper 525–C, p. C133–C136.

Kriss, A. E., 1940, Microorganism in permafrost subsoil: Mikrobiologiya, v. 9, p. 879–887.

Krueger, H. W., and Weeks, C. F., 1966, Geochron Laboratories, Inc. radiocarbon measurements 2: Radiocarbon v. 8, p. 142–160.

Kulp, J. L., Tryon, L. E., Eckelman, W. R., and Snell, W. A., 1952, Lamont natural radiocarbon measurements II: Science, v. 116, p. 409–414.

Kunkle, G. R., 1958, Multiple glaciation in the Jago River area, northeastern Alaska: Michigan Univ., M.S. thesis, 41 p.

Kurtén, Bjorn, 1968, Pleistocene mammals of Europe: Chicago, Aldine Publishing Co., 317 p.

LaChapelle, E. R., 1968, The character of snow avalancing induced by the Alaska earthquake, *in* The great Alaska earthquake of 1964 (Hydrology Volume): Natl. Acad. Sci. Pub. 1603, p. 355–361.

Lachenbruch, A. H., 1957, Thermal effects of the ocean on permafrost: Geol. Soc. America Bull., v. 68, p. 1515–1529.

——— 1960a, Thermal contraction cracks and ice wedges in permafrost, *in* Short papers in the geological sciences 1960: U.S. Geol. Survey Prof. Paper 400–B, p. B404–B406.

——— 1960b, Contraction-crack polygons, *in* Short papers in the geological sciences 1960: U.S. Geol. Survey Prof. Paper 400–B, p. B406–B409.

——— 1961, Depth and spacing of tension cracks: Jour. Geophys. Research, v. 66, p. 4273–4292.

——— 1962, Mechanics of thermal contraction cracks and ice-wedge polygons in permafrost: Geol. Soc. America Spec. Paper 70, 69 p.

——— 1966, Contraction theory of ice-wedge polygons—A qualitative discussion, *in* Permafrost Internat. Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 63–71.

——— 1968, Permafrost, *in* Fairbridge, R. W., ed., The encyclopedia of geomorphology: New York, Reinhold Pub. Corp., p. 833–839.

——— 1970a, Some estimates of the thermal effects of a heated pipeline in permafrost: U.S. Geol. Survey Circ. 632, 23 p.

——— 1970b, Thermal considerations in permafrost, *in* Adkison, W. L., and Brosgé, M. M., eds., Proceedings of the Seminar on the North Slope of Alaska: Los Angeles, Pacific Sec. Am. Assoc. Petroleum Geologists, p. J1–J5.

Lachenbruch, A. H., and Brewer, M. C., 1962, Geothermal evidence for recent climatic change near Barrow, Alaska [abs.]: Alaskan Sci. Conf., 12th, College, Alaska, 1961, Proc., p. 127; also Geol. Soc. America Spec. Paper 68, p. 117.

Lachenbruch, A. H., Brewer, M. C., Greene, G. W., and Marshall, B. V., 1962, Temperature in permafrost, *in* Temperature—its measurement and control in science and industry: New York, Reinhold Publishing Corp., v. 3, pt. 1, p. 791–803.

Lachenbruch, A. H., Greene, G. W., and Marshall, B. V., 1961, Interim report on geothermal studies, Ogotoruk Valley, northwestern Alaska, *in* Kachadoorian, Reuben, and others, 1961, Geological investigations in support of Project Chariot, Phase 3, in the vicinity of Cape Thompson, northwestern Alaska, preliminary report: U.S. Geol. Survey TEI–779, p. 68–81, issued by U.S. Atomic Energy Comm. Tech. Inf. Service, Oak Ridge, Tenn.

——— 1966, Permafrost and the geothermal regimes, *in* Wilimovsky, N.J., ed., Environment of the Cape Thompson region, Alaska: Oak Ridge, U.S. Atomic Energy Comm., p. 149–163.

Lachenbruch, A. H., and Marshall, B. V., 1969, Heat flow in the Arctic: Arctic, v. 22, no. 3, p. 300–311.

Lane, A. C., 1946, Northern climate variations affecting geothermal initial: Canadian Inst. Mining and Metallurgy Bull. 411, p. 397–402.

Larsen, Helge, 1951, De Dansk-Amerikanske, Alaska-ekspeditioner 1949–1950: Geog. Tiddskr., v. 51, p. 63–93 (English summary, p. 92–93).

——— 1953, Archaeological investigations in Alaska since 1939: Polar Rec., v. 6, p. 593–607.

——— 1968, Trail Creek: Acta Arctica, v. 15, 77 p.

Larsen, Helge, and Rainey, F. G., 1948, Ipiatak and the Arctic whale-hunting culture: Am. Mus. Nat. History, Anthropol. Papers, v. 42, 276 p.

Larsson, Peter, 1960, A preliminary investigation of the meteorological conditions on the Chamberlin Glacier, 1958: Arctic Inst. North America Research Paper 2, 89 p.

Laughlin, W. S., 1963, Eskimos and Aleuts; their origins and evolution: Science, v. 142, p. 633–645.

Lawrence, D. B., 1950, Glacier fluctuation for six centuries in southeastern Alaska and its relations to solar activity: Geog. Rev., v. 40, p. 191–223.

——— 1958, Glaciers and vegetation in southeastern Alaska: Am. Scientist, v. 46, p. 89–122.

Leffingwell, E. de K., 1915, Ground-ice wedges; the dominant form of ground-ice on the north coast of Alaska: Jour. Geology, v. 23, p. 635–654.

———1919, The Canning River region, northern Alaska: U.S. Geol. Survey Prof. Paper 109, 251 p.

Leonard, A. B., 1952, Illinoian and Wisconsin molluscan faunas in Kansas: Kansas Univ. Paleont. Contr. [No. 9, Mollusca], art. 4, 38 p.

Leonard, A. B., and Frye, D. G., 1960, Wisconsin molluscan faunas of the Illinois valley region: Illinois Geol. Survey Circ. 304, 32 p.

Leopold, L. B., 1951, Pleistocene climate in New Mexico: Am. Jour. Sci., v. 249, p. 152–168.

Lerbekmo, J. F., and Campbell, F. A., 1969, Distribution composition, and source of the White River Ash, Yukon Territory: Canadian Jour. Earth Sci., v. 6, p. 109–116.

Lerbekmo, J. F., Hanson, L. W., and Campbell, F. A., 1968, Application of particle size distribution to determination of source of a volcanic ash deposit: Internat. Geol. Cong., 23d, Prague 1968, Proc., v. 2, p. 283–295.

Lerbekmo, J. F., Westgate, J. A., Smith, D. G. W., and Denton, G. H., 1973, New data on the eruptive history of the White River vent, Alaska: Internat. Assoc. Quaternary Research, 9th Cong., New Zealand, 1973, Abs. Vol., p. 208.

Lewellen, R. I., 1973, The occurrence and characteristics of nearshore permafrost, northern Alaska, *in* Permafrost, The North American contribution to the Second International Conference, Yakutsk, Siberia, U.S.S.R.: Washington, D.C., Natl. Acad. Sci., p. 131–136.

Liapunova, E. A., and Vorontsov, N. N., 1970, Chromosomes and some issues of the evolution of the ground squirrel genus *Citellus* (Rodentia: Sciuridae): Experentia, v. 20, p. 1033–1038.

Liestøl, O., 1962, Errors and number of the glaciers in the snow fields, *in* Hoel, A., and Werenskiold, W., Glaciers and snow fields in Norway: Norsk Polar Inst. Skrister, no. 114, p. 28–54.

Lindholm, G. F., Thomas, L. A., Davidson, D. T., Handy, R. L., and Roy, C. J., 1959, Silt near Big Delta and Fairbanks, *in* Davidson, D. T., Roy, C. J., and others, 1959, The geology and engineering characteristics of some Alaskan soils: Iowa State Univ. Bull. 186, p. 33–70.

Lissey, Allan, 1970, Permafrost geohydrology (a) Pingos, Yukon, *in* the Federal Groundwater Program: Canada Dept. Energy, Mines and Resources Inland Waters Br. Rept. Ser. no. 9, p. 333–341.

Livingstone, D. A., 1954, On the orientation of lake basins (Alaska): Am. Jour. Sci., v. 252, p. 547–554.

——— 1955, Some pollen profiles from Arctic Alaska: Ecology, v. 36, p. 587–600.

——— 1957, Pollen analysis of a valley fill near Umiat, Alaska: Am. Jour. Sci., v. 255, p. 254–260.

Livingstone, D. A., Bryan, Kirk, Jr., and Leahy, R. G., 1958, Effect of an arctic environment on the origin and development of freshwater lakes: Limnology and Oceanography, v. 3, p. 192–214.

Long, Austin, and Mielke, J. E., 1967, Smithsonian Institution radiocarbon measurements 4: Radiocarbon, v. 9, p. 368–381.

Lotspeich, F. B., Mueller, E. W., and Frey, P. J., 1969, Some chemical and biological properties of a pingo lake in east-central Alaska: Federal Water Quality Admin., College, Alaska, Water Lab., no. 1, 12 p.

Lozinski, Walery, 1909, Uber die mechanische Verwitterung der Sandsteine im germassigten Klima: Acad. Sci. de Carcovie, Cl. des Sci., Math., et Nat., Bull., p. 1–25.

Lundquist, Jan, 1969, Earth and ice mounds; a terminological discussion, *in* Péwé, T. L., ed., The periglacial environment, past and present: Montreal, McGill-Queens Univ. Press, p. 203–215.

MacCarthy, G. R., 1958, Glacial boulders on the Arctic coast of Alaska: Arctic, v. 11, p. 71–85.

McCulloch, D. S., 1967, Quaternary geology of the Alaskan shore of Chukchi Sea, *in* Hopkins, D. M., ed., The Bering Land Bridge: Stanford, Stanford Univ. Press, p. 91–120.

McCulloch, D. S., and Hopkins, D. M., 1966, Evidence for an early Recent warm interval in northwest Alaska: Geol. Soc. America Bull., v. 77, p. 1089–1108.

McCulloch, D. S., Taylor, J. S., and Rubin, Meyer, 1965, Stratigraphy, non-marine mollusks, and radiometric dates from Quaternary deposits in the Kotzebue Sound area, western Alaska: Jour. Geology, v. 73, p. 442–453.

Mackay, J. R., 1963, The Mackenzie Delta area, N.W.T.: Canada Dept. Mines and Tech. Survey Geog. Br. Mem. 8, 202 p.

——— 1966, Pingos in Canada, *in* Permafrost Internat. Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 71–76.

——— 1968, Discussion of the theory of pingo formation by water expulsion in a region affected by subsidence: Jour. Glaciology, v. 7, p. 346–357.

——— 1971, The origin of massive icy beds in permafrost western Arctic coast, Canada: Canada Jour. Earth Sci., v. 8, no. 4, p. 397–422.

——— 1972a, Some observations on ice wedges, Garry Island, N.W.T., *in* Kerfoot, D. E., ed., Mackenzie Delta area monograph: Internat. Geog. Cong., 22d, St. Catharines, Ontario, Brock Univ., p. 131–139.

——— 1972b, The world of underground ice: Assoc. Am. Geographers Annals, v. 62, no. 1, p. 1–22.

——— 1973, The growth of pingos, western arctic coast, Canada: Canadian Jour. Earth Sci., v. 10, p. 979–1004.

Mackay, J. R., and Black, R. F., 1973, Origin, composition, and structure of perennially frozen ground and ground ice, *in* Permafrost, The North American contribution to the Second International Conference, Yakutsk, Siberia, U.S.S.R.: Washington, D.C., Natl. Acad. Sci., p. 185–192.

Mackay, J. R., Mathews, W. H., and MacNeish, R. S., 1961, Geology of the Engigstciak archeological site, Yukon Territory: Arctic, v. 14, p. 25–52.

Mackay, J. R., and Stager, J. K., 1966, The structure of some pingos in the Mackenzie Delta Area, N.W.T.: Geog. Bull., v. 18, no. 14, p. 360–368.

McKennan, R. A., and Cook, J. P., 1968, Pre-history of Healy Lake, Alaska: Internat. Cong. Anthropol. Ethnolocol. Sci., 8th, Tokyo, Sept. 1968.

McKenzie, G. D., 1970a, Some properties and age of volcanic ash in Glacier Bay National Monument: Arctic, v. 23, no. 1, p. 46–49.

——— 1970b, Glacial geology of Adams Inlet, southeastern Alaska: Inst. Polar Studies Rept. No. 25, 121 p.

McKenzie, G. D., and Goldthwait, R. P., 1971, Glacial history of the last eleven thousand years in Adams Inlet, southeastern Alaska: Geol. Soc. America Bull., v. 82, p. 1767–1782.

MacNeil, F. S., 1957, Cenozoic megafossils of northern Alaska: U.S. Geol. Survey Prof. Paper 294–C, p. 99–126.

——— 1967, Cenozoic pectinids of Alaska, Iceland, and other northern regions: U.S. Geol. Survey Prof. Paper 553, 57 p.

MacNeil, F. S., Mertie, J. B., and Pilsbry, H. A., 1943, Marine invertebrate faunas of the buried beaches near Nome, Alaska: Jour. Paleontology, v. 17, p. 69–96.

Maddren, A. G., 1905, Smithsonian exploration in Alaska in 1904, in search of mammoth and other fossil remains: Smithsonian Misc. Colln. 49, 117 p.

Manley, Gordon, 1949, The snowline in Britain: Geog. Annals, no. 31, p. 179–193.
——— 1959, The late-glacial climate of northwest England: Liverpool and Manchester Geol. Jour., v. 2, pt. 2, p. 188–215.
Marangunic, Cedomir, and Bull, Colin, 1968, The landslide on the Sherman Glacier, *in* The great Alaska earthquake of 1964 (Hydrology Volume): Natl. Acad. Sci. Pub. 1603, p. 383–394.
Marcus, M. G., 1968, Effects on glacier-dammed lakes in the Chugach and Kenai Mountains, *in* The great Alaska earthquake of 1964 (Hydrology Volume): Natl. Acad. Sci. Pub. 1603, p. 329–347.
Markov, K. K., and Popov, A. J., eds., 1959, Lednikovyy period na territorii Evropeyskoy chasti i Sibiri [The ice age in the European section of USSR and in Siberia]: Moscow, State Lomonosov Univ. of Moscow, 599 p.
Matthews, J. V., Jr., 1968a, A paleoenvironmental analysis of three Late Pleistocene Coleopterous assemblages from Fairbanks, Alaska: Alaska Univ., M.S. thesis, 59 p.
——— 1968b, A paleoenvironmental analaysis of three late Pleistocene Coleopterous assemblages from Fairbanks, Alaska: Quaestiones Entmologicae, v. 4, p. 202–224.
——— 1970, Quaternary environmental history of interior Alaska; pollen samples from organic colluvium and peats: Arctic and Alpine Research, v. 2, no. 4, p. 241–251.
——— 1974, Quaternary environments at Cape Deceit (Seward Peninsula, Alaska); evolution of a tundra ecosystem: Geol. Soc. America Bull., v. 85, p. 1353–1384.
Mayo, L. R., 1963, 1961 meteorology and mass balance of Gulkana Glacier, central Alaska Range, Alaska: Alaska Univ., M.S. thesis, 52 p.
Mayo, L. R., and Péwé, T. L., 1963, Ablation and net-total radiation, Gulkana Glacier, central Alaska Range, Alaska, *in* Kingery, W. D., ed., Ice and snow—properties, processes and applications, Massachusetts Inst. Technology, 1962, Proc.: Cambridge, Mass., M.I.T. Press, p. 633–643.
Meek, C. E., 1923, Notes on stratigraphy and Pleistocene fauna from Peard Bay, Arctic Alaska: California Univ., Dept. Geol. Sci. Bull., v. 14, no. 13, p. 409–422.
Meier, M. F., 1962, Proposed definitions for glacier mass budget terms: Jour. Glaciology, v. 4, no. 33, p. 252–261.
——— 1969, Seminar on the causes and mechanics of glacier surges, St. Hilaire, Canada, September 10–11, 1968, a summary: Canadian Jour. Earth Sci., v. 6, no. 4, p. 987–989.
Meier, M. F., and Post, A. S., 1962, Recent variations in mass net budgets of glaciers in western North America: Internat. Assoc. Sci. Hydrology Pub. 58, p. 63–77.
——— 1969, What are glacier surges?: Canadian Jour. Earth Sci., v. 6, no. 4, p. 807–817.
Meier, M. F., Tangborn, W. V., Mayo, L. R., and Post, Austin, 1971, Combined ice and water balances of Gulkana and Wolverine Glaciers, Alaska, and South Cascade Glacier, Washington, 1965 and 1966 hydrologic years: U.S. Geol. Survey Prof. Paper 715–A, 23 p.
Mendenhall, W. C., 1900, A reconnaissance from Resurrection Bay to Tanana River, Alaska, in 1898: U.S. Geol. Survey, 20th Ann. Rept., pt. 7, p. 265–340.
——— 1905, Geology of the central Copper River region, Alaska: U.S. Geol. Survey Prof. Paper 41, 133 p.
Mertie, J. B., Jr., 1925, Geology and gold placers of the Chandalar district, Alaska: U.S. Geol. Survey Bull. 773–E, p. 215–263.
——— 1937, The Yukon-Tanana region, Alaska: U.S. Geol. Survey Bull. 872, 276 p.
——— 1940, The Goodnews platinum deposits, Alaska: U.S. Geol. Survey Bull. 918, 97 p.
Mertie, J. B., Jr., and Harrington, G. L., 1924, The Ruby-Kuskokwim region, Alaska: U.S. Geol. Survey Bull. 754, 129 p.
Mielke, J. E., and Long, Austin, 1969, Smithsonian Institution radiocarbon measurements 5: Radiocarbon, v. 11, p. 163–182.
Miller, D. J., 1953a, Late Cenozoic marine glacial sediments and marine terraces of Middleton Island, Alaska: Jour. Geology, v. 61, p. 17–40.
——— 1953b, Geology and petroleum possibilities of Gulf of Alaska Tertiary Province [abs.]: Am. Assoc. Petroleum Geologists Bull., v. 37, p. 1132–1133.
——— 1957, Geology of the southeastern part of the Robinson Mountains, Yakataga district, Alaska: U.S. Geol. Survey Oil and Gas Inv. Map OM–187, scale 1:63,360.
——— 1961a, Geology of the Yakutat District, Gulf of Alaska Tertiary Province, Alaska: U.S. Geol. Survey open-file report.
——— 1961b, Stratigraphic occurrence of *Lituyapecten* in Alaska: U.S. Geol. Survey Prof. Paper 354–K, p. 241–249.
Miller, D. J., Payne, G. T., and Gryc, George, 1959, Geology of possible petroleum provinces in Alaska, with an annotated bibliography by E. H. Cobb: U.S. Geol. Survey Bull. 1094, 131 p.
Miller, M. M., 1952, ed., Scientific observations of the Juneau Icefield Research Project, Alaska, 1949 field season: J.I.R.P. Rept. No. 6, 162 p.
——— 1953, Juneau Icefield Research Project, 1951 winter season: J.I.R.P. Rept. No. 8, 51 p.
——— 1965a, Alaskan glacier variations and the implications of recent tectonic activity [abs.]: Geol. Soc. America Spec. Paper 82, p. 359.
——— 1965b, Some major effects of the 1964 Good Friday earthquake on Alaskan coastal glaciers [abs.]: Am. Geophys. Union Trans., v. 46, p. 79–80.
——— 1970, 1946–1962 Survey of the regional pattern of Alaskan glacier variations, *in* National Geographic Society research reports 1961–1962: Natl. Geog. Soc., p. 167–189.
Miller, R. D., and Dobrovolny, Ernest, 1959, Surficial geology of Anchorage and vicinity, Alaska: U.S. Geol. Survey Bull. 1093, 128 p.
Mitchell, J. M., Jr., 1965, Theoretical paleoclimatology, *in* Wright, H. E., Jr., and Frey, D. G., eds., The Quaternary of the United States: Princeton, Princeton Univ. Press, p. 881–901.
——— 1968, Concluding remarks, *in* Mitchell, J. M., Jr., ed., Causes of climatic change: Am. Meteoritical Soc., Meteorol. Monographs, v. 8, no. 30, p. 155–159.
Moffit, F. H., 1942, Geology of the Gerstle River district, Alaska, with a report on the Black Rapids Glacier: U.S. Geol. Survey Bull. 926–B, p. 107–160.
Moffit, F. H., and Capps, S. R., 1911, Geology and mineral resources of the Nizina district, Alaska: U.S. Geol. Survey Bull. 448, 108 p.
Moffit, F. H., and Knopf, Adolph, 1910, Mineral resources of the Nabesna-White River district, Alaska, with a section on the Quaternary by S. R. Capps: U.S. Geol. Survey Bull. 417, 64 p.
Moore, G. W., 1960, Recent eustatic sea-level fluctuations recorded by Arctic beach ridges, *in* Short papers in the geological sciences 1960: U.S. Geol. Survey Prof. Paper 400–B, p. B335–B337.
Moore, G. W., and Giddings, J. L., 1962, Record of 5000 years of Arctic wind direction recorded by Alaskan beach ridges: Geol. Soc. America Spec. Paper 68, p. 232.
Moores, E. A. 1962, Configuration of the surface velocity profile of Gulkana Glacier, central Alaska Range, Alaska: Alaska Univ., M.S. thesis, 47 p.
Mörner, N. A., 1971, The position of the ocean level during the interstadial at about 30,000 B. P.—a discussion from a climatic-glaciologic point of view: Canadian Jour. Earth Sci., v. 8, p. 132–143.
Morrison, R. B., 1969, The Pleistocene-Holocene boundary; an evaluation of the various criteria used for determining it on a provincial basis, and suggestions for establishing it world-wide: Geologie en Mijnbouw, v. 48, p. 363–371.
Moxham, R. M., and Eckhart, R. A., 1956, Marl deposits in the Knik Arm area, Alaska: U.S. Geol. Survey Bull. 1039–A, 23 p.

Mozley, Alan, 1937, Frozen ground in the Subarctic region and its biological significance: Scottish Geog. Mag., v. 53, p. 266–270.

Muir, John, 1884, On the glaciation of the Arctic and the Subarctic regions visited by the U.S. steamer *Corwin* in the year 1881, *in* Hopper, C. L., Report of the cruise of the U.S. Revenue steamer *Thomas Corwin* in the Arctic Ocean, 1881: U.S. 48th Cong., 1st sess., Senate Ex. Doc. 204, p. 135–147.

Muller, E. H., 1953, Northern Alaska Peninsula and eastern Kilbuck Mountains, Alaska, *in* Péwé, T. L., and others, Multiple glaciation in Alaska, a progress report: U.S. Geol. Survey Circ. 289, p. 2–3.

——— 1955, Bristol Bay region, *in* Hopkins, D. M., Karlstrom, T. N. V., and others, Permafrost and ground water in Alaska: U.S. Geol. Survey Prof. Paper 264–F, p. 131–133.

Muller, E. H., and Coulter, H. W., 1957, The Knife Creek Glaciers of Katmai National Monument, Alaska: Jour. Glaciology, v. 3, no. 22, p. 116–122.

Müller, Fritz, 1963, Observations on pingos [Translation by D. A. Sinclair of "Beobachtung uber—Pingos": Medd. om Gronland, v. 153, p. 1–127, 1959]: Natl. Research Council Canada Tech. Trans. 1073, 117 p.

Muller, S. W., 1945, Permafrost or permanently frozen ground and related engineering problems: U.S. Engineers Office, Strategic Eng. Study Spec. Rept. 62, 136 p. (Repr., 1947, Ann Arbor, Mich., J. W. Edwards Inc.).

Murie, O. J., 1951, The elk of North America: Harrisburg, The Stackpole Co., 376 p.

Nakagawa, Hisao, Niitsuma, Nobuaki, and Elmi, Carlo, 1971, Pliocene and Pleistocene magnetic stratigraphy in La Castella area, southern Italy—a preliminary report: Quaternary Research, v. 1, p. 360–368.

National Academy of Science–National Research Council, 1968, The Great Alaska Earthquake of 1964 (Hydrology Volume): Natl. Acad. Sci. Pub. 1603, 441 p. Human Ecology Volume, 1970, 510 p.; Biology Volume, 1971, 252 p.; Geology Volume, 1972, 831 p.; Seismology and Geodesy Volume, 1972, 596 p.; Oceanography and Coastal Engineering Volume, 1972, 556 p.; Engineering Volume 1973, 1190 p.

Nelson, Hans, and Hopkins, D. M., 1972, Sedimentary processes and distribution of particulate gold in the northern Bering Sea: U.S. Geol. Survey Prof. Paper 689, 27 p.

Nichols, D. R., 1956, Permafrost and ground-water conditions in the Glennallen area, Alaska: U.S. Geol. Survey open-file report 392, 18 p.

——— 1960, Slump structures in Pleistocene lake sediments, Copper River Basin, Alaska, *in* Short papers in the geological sciences 1960: U.S. Geol. Survey Prof. Paper 400–B, p. B353–B354.

——— 1961, Analyses of gas and water from two mineral springs in the Copper River Basin, Alaska, *in* Short papers in the geologic and hydrologic sciences 1961: U.S. Geol. Survey Prof. Paper 424–D, p. D191–D194.

——— 1963, Origin of the course of the Copper River, Alaska [abs.]: Geol. Soc. America Spec. Paper 73, p. 216.

——— 1966, Permafrost in the Recent Epoch, *in* Permafrost Internat. Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 172–175.

Nichols, D. R., and Yehle, L. A., 1961a, Mud volcanoes in the Copper River Basin, Alaska, *in* Raasch, G. O., ed., Geology of the Arctic: Toronto, Univ. Toronto Press, v. 2, p. 1063–1087.

——— 1961b, Highway construction and maintenance problems in permafrost regions, *in* Symposium on geology as applied to highway engineering, 12th, Ann., Knoxville 1961: Tennessee Univ. Eng. Expt. Sta. Bull. 24, p. 19–29.

——— 1969, Engineering geologic map of the southeastern Copper River basin, Alaska: U.S. Geol. Survey Misc. Inv. Map I–524, scale 1:125,000.

Nordenskiold, A. E., 1881, The voyage of the Vega round Asia and Europe: London, MacMillan & Co., v. 1, 524 p., v. 2, 464 p.

Nowak, Michael, 1969, Archeological dating by means of volcanic ash strata [abs.]: Dissert. Abs. Internat., Sec. B, Sci. and Eng., v. 29, no. 10, p. 3598B–3599B.

Osborn, H. F., 1942, Proboscidea: New York, Am. Mus. Nat. History, v. 2, p. 805–1075.

Ostenso, N. A., Sellmann, P. V., and Péwé, T. L., 1965, The bottom topography of Gulkana Glacier, Alaska Range, Alaska: Jour. Glaciology, v. 5, p. 651–660.

Ostrem, Gunnar, 1964, Ice-cored moraines in Scandinavia: Geog. Annals, v. 46, no. 3, p. 282–337.

——— 1966, The height of the glaciation limit in southern British Columbia and Alberta: Geog. Annals, v. 48, ser. A, no. 3, p. 126–138.

O'Sullivan, J. B., 1961, Quaternary geology of the Arctic coastal plain, northern Alaska: Iowa State Univ., Ph.D. thesis, 191 p.

——— 1966, Geochemistry of permafrost, Barrow, Alaska, *in* Permafrost Internat. Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 30–37.

Patten, D. W., 1966, The biblical flood and the ice epoch: Seattle, Pacific Meridian Pub. Co., 336 p.

Patton, W. W., Jr., and Hoare, J. M., 1968, The Kaltag fault, west-central, Alaska, *in* Geological Survey research 1968: U.S. Geol. Survey Prof. Paper 600–D, p. D147–D153.

Payne, T. G., compiler, 1955, Mesozoic and Cenozoic tectonic elements of Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–84, scale 1:5,000,000.

Peltier, L. C., 1950, The geographic cycle in periglacial regions as it is related to climatic geomorphology: Assoc. Am. Geographers Annals, v. 40, p. 214–236.

Péwé, T. L., 1947, Permafrost and geomorphology in the lower Yukon River Valley, Alaska [abs.]: Geol. Soc. America Bull., v. 58, p. 1256.

——— 1948, Terrain and permafrost of the Galena Air Base, Galena, Alaska: U.S. Geol. Survey Permafrost Program Progress Rept. 7, 52 p.

——— 1949, Preliminary report of permafrost investigations in the Dunbar area, Alaska: U.S. Geol. Survey Circ. 42, 3 p.

——— 1951a, An observation of wind-blown silt: Jour. Geology, v. 59, p. 399–401.

——— 1951b, Recent history of Black Rapids Glacier, Alaska [abs.]: Geol. Soc. America Bull., v. 62, p. 1558.

——— 1952a, Geomorphology of the Fairbanks area: Stanford Univ., Ph.D. thesis, 220 p.

——— 1952b, Preliminary report of multiple glaciation in the Big Delta area, Alaska [abs.]: Geol. Soc. America Bull., v. 63, p. 1289.

——— 1954, Effect of permafrost on cultivated fields, Fairbanks area, Alaska: U.S. Geol. Survey Bull. 989–F, p. 315–351.

——— 1955, Origin of the upland silt near Fairbanks, Alaska: Geol. Soc. America Bull., v. 66, p. 699–724.

——— 1957a, Preliminary report of the 1957 advance of Muldrow Glacier, Alaska [abs.]: Geol. Soc. America Bull., v. 68, p. 1908–1909.

——— 1957b, Recent history of Canwell and Castner Glaciers, Alaska [abs.]: Geol. Soc. America Bull., v. 68, p. 1779.

——— 1958a, Permafrost and its effect on life in the north, *in* Arctic Biology: Oregon State College, Ann. Biology Colloquium, 18th, Corvallis, p. 12–25.

——— 1958b, Geology of the Fairbanks (D–2) quadrangle, Alaska: U.S. Geol. Survey Geol. Quad. Map GQ–110, scale 1:63,360.

——— 1959, Sand wedge polygons (tesselations) in the McMurdo Sound region, Antarctica: Am. Jour. Sci., v. 257, p. 545–552.

——— 1961a, University of Alaska Gulkana Glacier expedition, 1960: Arctic, v. 14, p. 74–75.

——— 1961b, University of Alaska Gulkana Glacier expedition, 1961: Arctic, v. 14, p. 236–237.

——— 1961c, Multiple glaciation in the headwaters area of the Delta River, central Alaska, *in* Short papers in the geologic and hydrologic sciences 1961: U.S. Geol. Survey Prof. Paper 424–D, p. D200–D201.

——— 1962, Ice wedges in permafrost, lower Yukon River area near Galena, Alaska: Biul. Peryglacjalny, ser. 3, p. 65–76.

——— 1963, University of Alaska Gulkana Glacier project, 1962: Arctic, v. 16, p. 46–47.

——— 1964a, Good Friday earthquake, Alaska: Geotimes, v. 8, p. 9–10, 22–23.

——— 1964b, Preliminary geologic observations of March 27 earthquake in Alaska: Fairbanks Daily News Miner, v. 42, p. 21.

——— 1964c, Ice wedges in Alaska—classification, distribution, and climatic significance [abs.]: Geol. Soc. America Spec. Paper 76, p. 129.

——— 1965a, Fairbanks area, *in* Péwé, T. L., Ferrians, O. J., Jr., Nichols, D. R., and Karlstrom, T. N. V., Guidebook for Field Conf. F, Central and south central Alaska—Internat. Assoc. Quaternary Research, 7th Cong., USA 1965: Lincoln, Nebr., Nebraska Acad. Sci., p. 6–36.

——— 1965b, Middle Tanana River Valley, *in* Péwé, T. L., Ferrians, O. J., Jr., Nichols, D. R., and Karlstrom, T. N. V., Guidebook for Field Conf. F., Central and south central Alaska—Internat. Assoc. Quaternary Research, 7th Cong., USA 1965: Lincoln, Nebr., Nebraska Acad. Sci., p. 36–54.

——— 1965c, Delta River area, Alaska Range, *in* Péwé, T. L., Ferrians, O. J., Jr., Nichols, D. R., and Karlstrom, T. N. V., Guidebook for Field Conf. F, Central and south central Alaska —Internat. Assoc. Quaternary Research, 7th Cong., USA 1965: Lincoln, Nebr., Nebraska Acad. Sci., p. 55–93.

——— 1966a, Paleoclimatic significance of fossil ice wedges: Biul. Peryglacjalny, no. 15, p. 65–72.

——— 1966b, Ice-wedges in Alaska—classification, distribution and climatic significance, *in* Permafrost Internat. Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 76–81.

——— 1966c, Permafrost and its effect on life in the north: Corvallis, Oregon State Univ. Press, 40 p.

——— 1968a, Loess deposits of Alaska: Internat. Geol. Cong., 23d, Prague 1968, v. 8, p. 297–309.

——— 1968b, Introduction to avalanches on glaciers and the effects of snowslides, *in* The great Alaska earthquake of 1964 (Hydrology Volume): Natl. Acad. Sci. Pub. 1603, p. 351–354.

——— 1969a, ed., The periglacial environment; past and present: Montreal, McGill-Queens Univ. Press, 487 p.

——— 1969b, Terrasses d'altiplanation du Quaternaire anciene et moyen, Fairbanks, Alaska [abs.]: Internat. Assoc. Quaternary Research, 8th Cong., Resumés des commun., p. 41.

——— 1970a, Permafrost and vegetation on floodplains of subarctic rivers (Alaska), a summary, *in* Ecology of the subarctic regions—Proceedings of the Helsinki symposium: UNESCO, p. 141–142.

——— 1970b, Altiplanation terraces of early Quaternary age near Fairbanks, Alaska: Acta Geographica Lodziensia, no. 24, p. 357–363.

——— 1973, Ice wedge casts and past permafrost distribution in North America: Geoforum, v. 15, p. 15–26.

——— 1974a, Geomorphic processes in polar deserts, *in* Smiley, T. L., and Zumberge, J. H., eds., Polar deserts: Tucson, Univ. Arizona Press, p. 33–52.

——— 1974b, Permafrost: Encyclopedia Britannica, v. 14, p. 89–95.

——— 1975, Quaternary stratigraphic nomenclature in central Alaska: U.S. Geol. Survey Prof. Paper 862, 32 p.

Péwé, T. L., and Burbank, Lawrence, 1960, Multiple glaciation in the Yukon-Tanana Upland, Alaska; a photogeologic interpretation [abs.]: Geol. Soc. America Bull., v. 71, p. 2088.

Péwé, T. L., Burbank, Lawrence, and Mayo, L. R., 1967, Multiple glaciation in the Yukon-Tanana Upland, Alaska: U.S. Geol. Survey Misc. Inv. Map I–507, scale 1:500,000.

Péwé, T. L., Church, R. E., 1962, Age of the spit at Barrow, Alaska: Geol. Soc. America Bull., v. 73, p. 1287–1291.

Péwé, T. L., Church, R. E., and Andresen, M. J., 1969, Origin and paleoclimatic significance of large scale polygons in the Donnelly Dome area, Alaska: Geol. Soc. America Spec. Paper 109, 87 p.

Péwé, T. L., Ferrians, O. J., Jr., Nichols, D. R., and Karlstrom, T. N. V., 1965, Guidebook for Field Conf. F—Central and south central Alaska, Internat. Assoc. Quaternary Research, 7th Cong., USA 1965: Lincoln, Nebr., Nebraska Acad. Sci., 141 p.

Péwé, T. L., and Holmes, G. W., 1964, Geology of the Mt. Hayes D–4 quadrangle, Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–394, scale 1:63,360.

Péwé, T. L., and Hopkins, D. M., 1967, Mammal remains of pre-Wisconsin age in Alaska, *in* Hopkins, D. M., ed., The Bering Land Bridge: Stanford, Stanford Univ. Press, p. 266–270.

Péwé, T. L., Hopkins, D. M., and Giddings, J. L., Jr., 1965, Quaternary geology and archeology of Alaska, *in* Wright, H. E., Jr., and Frey, D. G., eds., The Quaternary of the United States: Princeton, Princeton Univ. Press, p. 355–374.

Péwé, T. L., and others, 1953, Multiple glaciation in Alaska, a progress report: U.S. Geol. Survey Circ. 289, 13 p.

Péwé, T. L., and Paige, R. A., 1963, Frost heaving of piles with an example from the Fairbanks area, Alaska: U.S. Geol. Survey Bull. 1111–I, p. 333–407.

Péwé, T. L., and Reger, R. D., 1972, Modern and Wisconsinan snowlines in Alaska: Internat. Geol. Cong., 24th, Montreal 1972, Proc., p. 187–197.

Péwé, T. L., and Rivard, N. R., 1961, Geologic map and section of the Fairbanks D–3 quadrangle, Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–340, scale 1:63,360.

Péwé, T. L., and Sellmann, P. V., 1971, Geochemistry of permafrost and Quaternary stratigraphy [abs.]: Geol. Soc. America Ann. Mtg., Absts. with Programs, v. 3, no. 7, p. 669.

Péwé, T. L., Wahrhaftig, Clyde, and Weber, F. R., 1966, Geologic map of the Fairbanks quadrangle, Alaska: U.S. Geol. Survey Misc. Inv. Map I–455, scale 1:250,000.

Plafker, George, 1956, Occurrence of diatomaceous earth near Kenai, Alaska: U.S. Geol. Survey Bull. 1039–B, p. 25–31.

——— 1965, Tectonic deformation associated with the 1964 Alaska earthquake: Science, v. 148, p. 1675–1687.

——— 1967, Geologic map of the Gulf of Alaska Tertiary Province, Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–484, scale 1:500,000.

——— 1968, Source areas of the Shattered Peak and Pyramid Peak landslides at Sherman Glacier, *in* The great Alaska earthquake of 1964 (Hydrology Volume): Natl. Acad. Sci. Pub. 1603, p. 374–382.

——— 1969, Tectonics of the March 27, 1964, Alaska earthquake: U.S. Geol. Survey Prof. Paper 543–I, 74 p.

——— 1971, Possible future petroleum resources of Pacific Margin Tertiary basin, Alaska: Am. Assoc. Petroleum Geol. Mem. 15, p. 120–135.

Plafker, George, and Miller, D. J., 1958, Glacial features and surficial deposits of the Malaspina district, Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–271, scale 1:125,000.

Plafker, George, and Rubin, Meyer, 1967, Vertical tectonic displacements in south-central Alaska during and prior to the Great 1964 Earthquake: Jour. Geoscience, v. 10, p. 53–66.

Popov, A. I., 1955, Origin and development of thick fossil ice, *in* Materialy k osnovam ucheniya o merzlykh zonakh zemnoy kory [The materials for the fundamentals of the study on frozen zones of the earth's crust]: Akad. Nauk SSSR, Inst. Merzlotovedeniya Trudy, no. 2, p. 5–25.

——— 1965, Podzemnyy led [Underground ice]: Moscow, Moscow Univ. Press, v. 1–3, 215 p., 179 p., and 175 p.

Porsild, A. E., 1938, Earth mounds in unglaciated Arctic northwestern America: Geog. Rev., v. 28, p. 46–58.

Porter, S. C., 1962, Geology of Anaktuvuk Pass, central Brooks Range, Alaska: Arctic Inst. North America Final Rept., 276 p.

——— 1964, Late Pleistocene glacial chronology of north-central Brooks Range, Alaska: Am. Jour. Sci., v. 262, p. 446–460.

——— 1966, Geology of the Anaktuvuk Pass, central Brooks Range, Alaska: Arctic Inst. North America Tech. Paper 18, 100 p.

——— 1967, Glaciation of Chagvan Bay area, southwestern Alaska: Arctic, v. 20, p. 227–246.

——— 1968, Determination of equilibrium-line altitudes for late Quaternary alpine glaciers: Geol. Soc. America Ann. Mtg., Abs. with Programs, p. 242.

——— 1970, Quaternary glacial record in Swat Kohistan, West Pakistan: Geol. Soc. America Bull., v. 81, no. 5, p. 1421–1446.

Porter, S. C., and Denton, G. H., 1967, Chronology of neoglaciation in the North American Cordillera: Am. Jour. Sci., v. 265, p. 177–210.

Post, A. S., 1960, The exceptional advances of Muldrow, Black Rapids, and Susitna Glaciers: Jour. Geophys. Research, v. 65, p. 3703–3712.

——— 1967, Effects of the March 1964 Alaska earthquake on glaciers: U.S. Geol. Survey Prof. Paper 544–D, 42 p.

——— 1969, Distribution of surging glaciers in western North America: Jour. Glaciology, v. 8, no. 53, p. 229–240.

Powers, H. A., Coats, R. R., and Nelson, W. H., 1960, Geology and submarine physiography of Amchitka Island, Alaska: U.S. Geol. Survey Bull. 1028–P, p. 521–554.

Prindle, L. M., 1908, The Fairbanks and Rampart quadrangles, Yukon-Tanana region, Alaska: U.S. Geol. Survey Bull. 337, 102 p.

——— 1913, A geological reconnaissance of the Fairbanks quadrangle, Alaska; with a detailed description of the Fairbanks district, by L. M. Prindle and F. J. Katz, and an account of lode mining near Fairbanks, by P. S. Smith: U.S. Geol. Survey Bull. 525, 220 p.

Quackenbush, L. S., 1909, Notes on Alaskan mammoth expedition of 1907 and 1908: Am. Mus. Nat. History Bull., v. 26, p. 87–130.

Ragan, D. M., 1964, Structures at the base of an ice fall on the Gulkana Glacier, eastern Alaska Range [abs.]: Geol. Soc. America Spec. Paper 76, p. 220–221.

——— 1966, An origin of foliation in glacial ice by shear, eastern Alaska Range [abs.]: Geol. Soc. America Spec. Paper 87, p. 132.

——— 1967, Planar and layered structures in glacial ice: Jour. Glaciology, v. 6, p. 565–567.

——— 1969, Structures at the base of an ice fall: Jour. Geology, v. 77, no. 6, p. 647–667.

Ragle, R. H., Sater, J. E., and Field, W. O., 1965a, Effects of the 1964 Alaskan earthquake on glaciers and related features: Arctic Inst. North America Research Paper 32, 44 p.

——— 1965b, Effects of the 1964 Alaskan earthquake on glaciers and related features: Arctic, v. 18, p. 135–137.

Rainey, F. G., 1939, Archaeology in central Alaska: Am. Mus. Nat. History, Anthropol. Papers, v. 36, p. 355–405.

——— 1941, Eskimo prehistory; the Okvik site on the Punuk Islands: Am. Mus. Nat. History, Anthropol. Papers, v. 37, 601 p.

Rampton, Vern, 1970, Neoglacial fluctuations of the Natazhat and Klutlan glaciers, Yukon Territory, Canada: Canadian Jour. Earth Sci., v. 7, no. 5, p. 1236–1263.

——— 1971, Late Quaternary vegetational and climatic history of the Snag Klutlan area, southwestern Yukon Territory, Canada: Geol. Soc. America Bull., v. 82, no. 4, p. 959–978.

Rapp, Anders, and Annesten, Lennart, 1969, Permafrost and tundra polygons in northern Sewden, *in* Péwé, T. L., ed., The periglacial environment—past and present: Montreal, McGill Univ. Press, p. 65–91.

Rausch, Robert, 1953, On the status of some Arctic mammals: Arctic, v. 6, p. 91–148.

Ray, C. E., 1971, Polar bear and mammoth on the Pribilof Islands: Arctic, v. 24, no. 1, p. 9–18.

Redozubov, D. B., 1946, Thermal field laws of the permafrost in the Vorkuta region: Akad. Nauk SSSR Inst. Merzlotovedeniia imeni. V. A. Obrucheva, v. 1 (1946), trans. by Leonid Stambulsheik and Jaroslav Jan Peel (1954), SIPRE Trans. 17, 20 p.

Reed, B. L., 1968, Geology of the Lake Peters area northeastern Brooks Range, Alaska: U.S. Geol. Survey Bull. 1236, 132 p.

Reed, J. C., 1969, The story of the Naval Arctic Research Laboratory: Arctic, v. 22, no. 3, p. 177–183.

Reed, J. C., and Coats, R. R., 1942, Geology and ore deposits of the Chicagof mining district, Alaska: U.S. Geol. Survey Bull. 929, 148 p.

Reed, J. C., Jr., 1960, Geology of the Mt. McKinley quadrangle, Alaska: U.S. Geol. Survey open-file report., 36 p.

——— 1961, Geology of the Mount McKinley quadrangle, Alaska: U.S. Geol. Survey Bull. 1108–A, 36 p.

Reger, R. D., 1964, Recent glacial history of Gulkana and College Glaciers, central Alaska Range, Alaska: Alaska Univ., M.S. thesis, 75 p.

——— 1968, Recent history of Gulkana and College Glaciers, central Alaska Range, Alaska: Jour. Geology, v. 76, p. 2–16.

Reger, R. D., and Péwé, T. L., 1969, Lichenometric dating in the central Alaska Range, *in* Péwé, T. L., ed., The periglacial environment; past and present: Montreal, McGill-Queens Univ. Press, p. 223–247.

Reger, R. D., Péwé, T. L., Hadleigh-West, Frederick, and Skarland, Ivar, 1964, Geology and archaeology of the Yardang Flint Station: Alaska Univ., Anthropol. Papers, v. 12, p. 92–100.

Reid, H. F., 1892, Studies of Muir Glacier, Alaska: Natl. Geog. Mag., v. 4, p. 19–55.

——— 1896, Glacier Bay and its glaciers (Alaska). U.S. Geol. Survey Ann. Rept. 16, pt. 1, p. 415–461.

Reid, J. R., Jr., 1968, Origin and characteristics of the superglacial drift, Martin River and Sioux glaciers, Alaska [abs.]: Geol. Soc. America Spec. Paper 101, p. 173.

——— 1970a, Late Wisconsin and Neoglacial history of the Martin River Glacier, Alaska: Geol. Soc. America Bull., v. 81, p. 3593–3604.

——— 1970b, Geomorphology and glacial geology of the Martin River Glacier, Alaska: Arctic, v. 23, no. 4, p. 254–267.

Reimnitz, Erk, and Von Huene, Roland, 1969, Bathmetry and isopach map of stratified Holocene sediments of Nuka Bay, Alaska: U.S. Geol. Survey open-file report, 3 sheets.

Repenning, C. A., Hopkins, D. M., and Rubin, Meyer, 1964, Tundra rodents in a late Pleistocene fauna from the Tofty district, central Alaska: Arctic, v. 17, p. 177–197.

Richter, D. H., and Matson, N. A., Jr., 1971, Quaternary faulting in the eastern Alaska Range: Geol. Soc. America Bull., v. 82, p. 1529–1540.

Richmond, G. M., 1965, Glaciation of the Rocky Mountains, *in* Wright, H. E., Jr., and Frey, D. G., eds., The Quaternary of the United States: Princeton, Princeton Univ. Press, p. 217–230.

Rickert, D. A., and Tedrow, J. C. F., 1967, Pedologic investigations of some aeolian deposits of northern Alaska: Soil Sci., v. 104, p. 250–262.

Ritchie, J. C., and Hare, F. K., 1971, Late-Quaternary vegetation and climate near the Arctic tree line of northwestern North America: Quaternary Research, v. 1, no. 3, p. 331–342.

Romer, Eugenjusz, 1929, A few remarks on the tree and névé-lines in the Canadian and Alaskan Cordillera: Przegląd Geog., v. 9, p. 227–251.

Rosenfeld, G. A., and Hussey, K. M., 1958, A consideration of the problem of oriented lakes: Iowa Acad. Sci. Proc., v. 65, p. 279–287.

Rubin, Meyer, and Alexander, Corrine, 1958, U.S. Geological Survey radiocarbon dates 4: Science, v. 127, p. 1476–1487.

——— 1960, U.S. Geological Survey radiocarbon dates 5: Am. Jour. Sci., Radiocarbon Supp., v. 2, p. 129–185.

Russell, I. C., 1892a, Mount St. Elias and its glaciers: Am. Jour. Sci., v. 43, p. 169–182.

——— 1892b, Origin of the gravel deposits beneath Muir Glacier, Alaska: Am. Geologist, v. 9, p. 190–197.

——— 1893, Malaspina Glacier, Alaska: Jour. Geology, v. 1, p. 219–245.

Rutter, N. W., 1965, Foliation pattern of Gulkana Glacier, Alaska Range, Alaska: Jour. Glaciology, v. 5, p. 711–718.

Sable, E. G., 1961, Recent recession and thinning of Okpilak Glacier, northeastern Alaska: Arctic, v. 14, p. 176–187.

——— 1965, Geology of the Romanzof Mountains Brooks Range northeastern Alaska: U.S. Geol. Survey, open-file report, 218 p.

Sackett, W. M., 1958, Ionium-uranium ratios in marine-deposited calcium carbonates and related materials: Washington Univ., Ph.D. thesis, 106 p.

Sainsbury, C. L., 1967a, Quaternary geology of western Seward Peninsula, Alaska, *in* Hopkins, D. M., ed., The Bering Land Bridge: Stanford, Stanford Univ. Press, p. 121–143.

——— 1967b, Upper Pleistocene features in the Bering Strait area, *in* Geological Survey research 1967: U.S. Geol. Survey Prof. Paper 575–D, p. D203–D213.

Sainsbury, C. L., Kachadoorian, Reuben, Campbell, R. H., and Scholl, D. W., 1965, Marine platform of probable Sangamon age, and associated terrace deposits, Cape Thompson area, northwestern Alaska: Arctic, v. 18, p. 230–245.

Sanderson, I. T., 1960, Riddle of the quick frozen giants: Saturday Evening Post, Jan. 16, p. 82.

Sater, J. E., 1959, Glacier studies of the McCall Glacier, Alaska: Arctic, v. 12, p. 82–86.

——— 1964, Effects on glaciers of Alaskan earthquakes of March 1964: Arctic, v. 17, p. 211.

Schenk, Erwin, 1965, Pleistozane Eiskeil—Polygonnetze im Vergleich mit heutzeitlichen Vorkommen in Alaska: Natur u. Mus., v. 95, p. 8–16.

——— 1966, Origin of string bogs, *in* Permafrost Internat. Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 155–159.

Schmidt, R. A. M., 1963, Pleistocene marine microfauna in the Bootlegger Cove Clay, Anchorage, Alaska: Science, v. 141, p. 350–351.

Schmidt, R. A. M., and Sellmann, P. V., 1966, Mummified Pleistocene ostracods in Alaska: Science, v. 153, p. 167–168.

Schmoll, H. R., Szabo, B. J., Rubin, Meyer, and Dobrovolny, Ernest, 1972, Radiometric dating of marine shells from the Bootlegger Cove Clay, Anchorage area, Alaska: Geol. Soc. America Bull., v. 83, no. 4, p. 1107–1114.

Scholl, D. W., Alpha, T. R., Marlow, M. S., and Buffington, E. C., 1974, Base map of the Aleutian–Bering Sea region: U.S. Geol. Survey Misc. Inv. Map I–879, scale 1:2,500,000.

Scholl, D. W., Buffington, E. C., and Hopkins, D. M., 1968, Geologic history of the continental margin of North America in the Bering Sea: Marine Geology, v. 6, p. 297–330.

Scholl, D. W., Creager, J. S., and others, 1971, Deep Sea Drilling Project Leg 19: Geotimes, v. 16, no. 11, p. 12–15.

Schrader, F. C., 1904, A reconnaissance in northern Alaska: U.S. Geol. Survey Prof. Paper 20, 139 p.

Schrader, F. C., and Spencer, A. C., 1901, The geology and mineral resources of a portion of the Copper River district, Alaska: U.S. Geol. Survey special publication, 94 p.

Schwatka, Frederick, 1885, Along Alaska's great river: New York, Cassell & Co., Ltd., 360 p.

Schweger, C. E., 1971, Late-Quaternary paleoecology of the Onion Portage region, northwestern Alaska [abs.]: Geol. Soc. America, 24th Ann. Mtg., Abs. with Programs, v. 3, no. 6, p. 413.

Scruton, P. C., 1953, Marine geology of the Near Island Shelf, Alaska: Scripps Inst. Oceanography, La Jolla, Calif., Ph.D. thesis, (U.S. Geol. Survey open-file report), 200 p.

Sears, P. B., 1935, Glacial and post-glacial vegetation: Bot. Rev., v. 1, p. 37–51.

——— 1938, Climatic interpretation of post-glacial pollen deposits in North America: Am. Meteorol. Soc. Bull., v. 19, p. 177–181.

Selli, Raimondo, 1967, The Pliocene-Pleistocene boundary in Italian marine sections and its relationship to continental stratigraphies, *in* Sears, Mary, ed., Progress in Oceanography, v. 4,: New York, Pergamon Press, p. 67–86.

Sellmann, P. V., 1962, Flow and ablation of Gulkana Glacier, Alaska: Alaska Univ., M.S. thesis, 36 p.

——— 1967, Geology of the USA CRREL permafrost tunnel, Fairbanks, Alaska: U.S. Army Materiel Command Cold Regions Research and Eng. Lab. Tech. Rept. 199, 33 p.

——— 1968, Geochemistry and ground ice structures; an aid in interpreting a Pleistocene section, Alaska [abs.]: Geol. Soc. America Spec. Paper 101, p. 197–198.

——— 1972, Additional information on the geology and properties of materials exposed in the USA CRREL permafrost tunnel: Hanover, N.H., U.S. Army Materiel Command Cold Regions Research and Eng. Lab. Spec. Rept., 16 p.

Sellmann, P. V., and Brown, Jerry, 1963, Near-surface lithology of the Barrow, Alaska area, a preliminary report: Alaskan Sci. Conf., 14th, Anchorage, Proc., p. 231–232.

——— 1965, Coring of frozen ground, Barrow, Alaska: U.S. Army Materiel Command Cold Regions Research and Eng. Lab. Spec. Rept. 81, 8 p.

Sellmann, P. V., Brown, Jerry, and Lewellen, R. I., 1964, Near-surface stratigraphy, Barrow, Alaska; core analyses [abs.]: Alaskan Sci. Conf., 15th, College, Proc., p. 98.

Sellmann, P. V., Brown, Jerry, and Schmidt, R. A. M., 1965, Late-Pleistocene stratigraphy, Barrow, Alaska [abs.]: Internat. Assoc. Quaternary Research, 8th Cong., Paris, Abstract Volume, p. 419–420.

Shannon and Wilson, Inc., 1964, Report on Anchorage area soil studies, Alaska to U.S. Army Engineer District, Anchorage, Alaska: Seattle, Wash., 109 p., app. A–K, pls., illus.

Sharp, R. P., 1941, Geology of the Wolf Creek area, St. Elias Range, Yukon Territory, Canada: Geol. Soc. America Bull., v. 54, p. 625–649.

——— Soil structures in the St. Elias Range, Yukon Territory: Jour. Geomorphology, v. 5, p. 274–301.

——— 1951a, Glacial history of Wolf Creek, St. Elias Range, Canada: Jour. Geology, v. 59, p. 97–117.

——— 1951b, Accumulation and ablation on the Seward-Malaspina glacier system, Canada-Alaska: Geol. Soc. America Bull., v. 62, p. 725–743.

——— 1956, Last major advance of Malaspina Glacier, Alaska [abs.]: Geol. Soc. America Bull., v. 67, p. 1782.

——— 1957, Structures within the Malaspina Glacier, Alaska [abs.]: Geol. Soc. America Bull., v. 68, p. 1845–1846.

Shearer, J. M., MacNab, R. F., Pelletier, B. R., and Smith, T. B., 1971, Submarine pingos in the Beaufort Sea: Science, v. 174, p.

816–818.

Sher, A. V., 1967, Saiga remains mined in the northeast of Siberia and in Alaska: Akad. Nauk SSSR Kom. Izucheniyu Chetvertich. Perioda Byull. 33, p. 97–112. (Translated from Russian by L. A. Hutchison.)

——— 1969, Early Pleistocene mammals of extreme northeastern Asia and their environment [abs.]: Internat. Assoc. Quaternary Research, 8th Cong., Paris, p. 135.

——— 1971, Mammals and stratigraphy of the Pleistocene of the extreme northeast of the U.S.S.R. and North America: Nauka, Moscow, 310 p.

Sheth, Madhusudan, 1971, A heavy mineral study of Pleistocene and Holocene sediments near Nome, Alaska: U.S. Geol. Survey open-file report, 83 p.

Shreve, R. L., 1966, Sherman landslide, Alaska: Science, v. 154, p. 1639–1643.

——— 1968, Novel features of the Sherman landslide, Alaska [abs.]: Geol. Soc. America Spec. Paper 101, p. 200.

Shumskiy, P. A., Krenke, A. N., and Zotikov, I. A., 1964, Ice and its changes, *in* Solid earth and interface phenomena, v. 2, Research in geophysics: Cambridge, M.I.T. Press, p. 425–460.

Shumskiy, P. A., Schvetzov, P. F., and Dostovolov, B. N., 1955, Ice veins, *in* Special applications of engineering geology in reconnaissance of underground veins in the regions of their occurrence: U.S.S.R. Acad. Sci. Pub., p. 12–32.

Shumskiy, P. A., and Vtyurin, B. I., 1966, Underground ice, *in* Permafrost Internat. Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 108–113.

Sigafoos, R. S., 1958, Vegetation of northwestern North America, as an aid in interpretation of geologic data: U.S. Geol. Survey Bull. 1061–E, p. 165–185.

Sigafoos, R. S., and Hopkins, D. M., 1952, Soil instability on slopes in regions of perennially-frozen ground (Alaska): Highway Research Board Spec. Rept. 2, p. 176–192.

Skarland, Ivar, 1949, The geography of Alaska in Pleistocene and early post-glacial time; a study of the environment from an anthropological viewpoint: Harvard Univ., Ph.D. thesis, 192 p.

Skinner, M. F., and Kaisen, O. C., 1947, The fossil *Bison* of Alaska and preliminary revision of the genus: Am. Mus. Nat. History Bull., v. 89, p. 123–256.

Smith, D. G. W., and Westgate, J. A., 1969, Electron probe technique for characterizing pyroclastic deposits: Earth and Planetary Sci. Letters, v. 5, p. 313–319.

Smith, D. G. W., Westgate, J. A., and Tomlinson, M. C., 1969, Characterization of pyroclastic units—a stratigraphic application of the microprobe: Electron Microprobe Analysis Natl. Conf., 4th, Pasadena, Calif., Proc., Paper 34, 3 p.

Smith, H. T. U., 1949, Physical effects of Pleistocene climate changes in nonglaciated areas—Eolian phenomena, frost action and stream terracing: Geol. Soc. America Bull., v. 60, p. 1485–1515.

Smith, P. S., and Mertie, J. B., 1930, Geology and mineral resources of northwestern Alaska: U.S. Geol. Survey Bull. 815, 351 p.

Smith, R. K., 1970, Late glacial Foraminifera from southeast Alaska and British Columbia and a world-wide high northern latitude shallow-water faunal province: Archives Sci., v. 23, no. 3, p. 675–701.

Smith, W. R., 1925, Aniakchak Crater, Alaska Peninsula: U.S. Geol. Survey Prof. Paper 132, p. 139–145.

Spetzman, L. A., 1959, Vegetation of the Arctic Slope of Alaska: U.S. Geol. Survey Prof. Paper 302–B, p. B19–B58.

Spurr, J. E., 1898, Geology of the Yukon gold district, Alaska: U.S. Geol. Survey 18th Ann. Rept., pt. 3, p. 87–392.

——— 1900, A reconnaissance in southwestern Alaska in 1898: U.S. Geol. Survey 20th Ann. Rept., pt. 7, p. 31–264.

Steuerwald, B. A., Clark, D. L., and Andrew, J. A., 1968, Magnetic stratigraphy of faunal patterns in Arctic Ocean sediments: Earth and Planetary Sci. Letters, v. 5, p. 79–85.

Stock, Chester, 1942, A ground sloth in Alaska: Science, v. 95, p. 552–553.

Stoneley, R., 1970, Discussion, *in* Lachenbruch, A. H., Thermal consideration in permafrost, *in* Adkison, W. L., and Brosgé, M. M., eds., Proceedings of the Seminar on the North Slope of Alaska: Los Angeles, Pacific Sec. Am. Assoc. Petroleum Geologists,, p. J3–J4.

Stout, J. H., and Brady, J. B., 1972, Quaternary movement on the Denali fault in the Delta River area, Alaska [abs.]: Geol. Soc. America, Cordilleran Sec., 68th Ann. Mtg., Abs. with Programs, v. 4, no. 3, p. 243.

Stout, J. H., Brady, J. D., Weber, F. R., and Page, R. A., 1972, Evidence for Quaternary movement on the McKinley strand of the Denali fault in the Delta River area, Alaska: Geol. Soc. America Bull., v. 84, no. 3, p. 939–947.

Stuckenrath, R., Jr., and Mielke, J. E., 1970, Smithsonian Institution radiocarbon measurements 6: Radiocarbon, v. 12, p. 193–204.

——— 1973, Smithsonian Institution radiocarbon measurements 8: Radiocarbon, v. 15, p. 388–424.

Stuiver, Minze, 1969, Yale natural radiocarbon measurements 9: Radiocarbon, v. 11, p. 545–658.

Stuiver, Minze, Armstrong, R. L., and Denton, G. H., 1969, Quaternary glacial chronology, White River Valley, northern Wrangell and St. Elias Mountains, Alaska: Geol. Soc. America, Abs. with Programs for 1969, pt. 7, p. 218.

Stuiver, Minze, Borns, H. W., and Denton, G. H., 1964, Age of a widespread layer of volcanic ash in the southwestern Yukon Territory: Arctic, v. 17, p. 259–260.

Stump, R. W., Handy, R. L., Davidson, D. T., and Thomas, L. A., 1959, Silt deposit in the Matanuska Valley, *in* Davidson, D. T., Roy, C. J., and others, The geology and engineering characteristics of some Alaskan soils: Iowa State Univ. Bull. 186, p. 3–32.

Suess, H. F., 1954, U.S. Geological Survey radiocarbon dates 1: Science, v. 120, p. 467–473.

Sumgin, M. I., 1927, Dalnoe Vostochnoe Geofizicheskoye Obchestvo [Perennially frozen soil in the limits of U.S.S.R.]: Far Eastern Geophys. Obs., Vladivostok, S.S.S.R., 372 p. (in Russian).

Sumgin, M. I., and others, 1940, Obshchee merzlotovedenie [General geocryology]: Moscow-Leningrad, Akad. Nauk S.S.S.R., 340 p. (in Russian).

Swain, F. M., 1961, Ostracoda from the Pleistocene Gubik Formation, Arctic Coastal Plain, Alaska, *in* Raasch, G. O., ed., Geology of the Arctic, v. 1: Toronto, Univ. Toronto Press, p. 600–606.

Swanston, D. W., 1969, A late-Pleistocene glacial sequence from Prince of Wales Island, Alaska: Arctic, v. 22, p. 25–33.

Taber, Stephen, 1929, Frost heaving: Jour. Geology, v. 37, p. 428–461.

——— 1930, The mechanics of frost heaving: Jour. Geology, v. 38, p. 303–317.

——— 1943, Perennially frozen ground in Alaska; its origin and history: Geol. Soc. America Bull., v. 54, p. 1433–1548.

——— 1953, Origin of Alaska silts: Am. Jour. Sci., v. 251, p. 321–336.

——— 1958, Complex origin of silts in the vicinity of Fairbanks, Alaska: Geol. Soc. America Bull., v. 69, p. 131–136.

Tagg, A. R., and Greene, H. G., 1970, High-resolution seismic survey of an offshore area near Nome, Alaska: U.S. Geol. Survey open-file report, 42 p.

Taliaferro, N. L., 1932, Geology of the Yakataga, Katalla, and Nichawak districts, Alaska: Geol. Soc. America Bull., v. 43, p. 749–782.

Tarr, R. S., and Martin, Lawrence, 1914, Alaskan glacier studies: Wash., D. C., Natl. Geog. Soc., 498 p.

Tedrow, J. C. F., 1969, Thaw lakes, thaw sinks and soils in northern Alaska: Biul. Peryglacjalny, no. 20, p. 337–344.

Tedrow, J. C. F., and Walton, G. F., 1964, Some Quaternary events of northern Alaska: Arctic, v. 17, p. 268–271.

Thiel, E. C., LaChapelle, E. R., and Behrendt, J. C., 1957, The thickness of Lemon Creek Glacier, Alaska as determined by gravity measurements by E. Thiel, E. LaChapelle, and J. Behrendt: Am. Geophys. Union Trans., v. 38, p. 745–749.

Thorarinsson, Sigurdur, 1944, Tefrokronologiska studier pa Island: Geog. Annaler, v. 26, p. 1–203 (English summary, p. 204–215).

——— 1949, Some tephrochronological contributions to the volcanology and glaciology of Iceland: Geog. Annaler, v. 31, p. 239–256.

Tikhomirov, B. A., 1958, Natural conditions and vegetation in the Mammoth epoch in northern Siberia: Problems of the North, no. 1, p. 168–188.

Tolmachoff, I. D., 1929, The carcasses of the mammoth and rhinoceros found in the frozen ground of Siberia: Am. Philos. Soc., v. 23, 74 p.

Trainer, F. W., 1960, Geology and ground-water resources of the Matanuska Valley agricultural area, Alaska: U.S. Geol. Survey Water-Supply Paper 1494, 116 p.

——— 1961, Eolian deposits of the Matanuska Valley agricultural area, Alaska: U.S. Geol. Survey Bull. 1121–C, 35 p.

Trainer, F. W., and Waller, R. M., 1965, Subsurface stratigraphy of glacial drift at Anchorage, Alaska, *in* Geological Survey research 1965: U.S. Geol. Survey Prof. Paper 525–D, p. D167–D174.

Tricart, Jean, and Cailleux, André, 1967, Traité de Géomorphologie, v. 2—Le modele des régions périglaciaires: Paris, Société d'Édition d'Enseignement Supérieur, 512 p.

Troll, Carl, 1944, Strukturboden, Solifluktion and Frostklimate der Erde: Geol. Rundschau, band 34, p. 545–695.

——— 1948, Der subnivale oder periglaziale Zyklus der denudation: Erdkunde, band 2, p. 1–21.

Tuck, Ralph, 1938, The loess of the Matanuska Valley, Alaska: Jour. Geology, v. 46, p. 647–653.

——— 1940, Origin of muck-silt deposits at Fairbanks, Alaska: Geol. Soc. America Bull., v. 51, p. 1295–1310.

Tuthill, S. J., 1966, Earthquake origin of superglacial drift on the glaciers of the Martin River area, south-central Alaska: Jour. Glaciology, v. 6, p. 83–88.

——— 1970, A comparison of the late Wisconsinan molluscan fauna of the Missouri Coteau district (North Dakota) with a modern Alaskan analogue [abs.]: Dissert. Abs. Internat., Sec. B, Sci. and Eng., v. 30, no. 9, p. 4211B.

Tuthill, S. J., Field, W. O., and Clayton, Lee, 1968, Postearthquake studies at Sherman and Sheridan Glaciers, *in* The great Alaska earthquake of 1964 (Hydrology Volume): Natl. Acad. Sci. Pub. 1603, p. 318–328.

Twenhofel, W. S., 1952, Recent shore-line changes along the Pacific coast of Alaska: Am. Jour. Sci., v. 250, p. 523–548.

UNESCO/IASH, 1970, Seasonal snow cover: UNESCO/IASH Tech. Paper in Hydrology 2, 38 p.

U.S. Army Corps of Engineers, 1946, Airfield pavement design—frost conditions, *in* An interim engineering manual for War Department construction: U.S. Army Corps Engineers, pt. 12, ch. 4, 10 p.

——— 1947, Report on frost investigation 1944–1945: U.S. Army Corps Engineers, 66 p.

——— 1949, Addendum No. 1, 1945–1947, report on frost investigation 1944–1945: U.S. Army Corps Engineers, 50 p.

——— 1954, Arctic and subarctic construction building foundations, *in* Engineering manual for War Department construction: U.S. Army Corps Engineers, pt. 15, ch. 4, 14 p.

——— 1956, Engineering problems and construction in permafrost regions, *in* The dynamic north: U.S. Office Naval Operations [Polar Projects] Bk. 2 (OPNAY PO3–17), 53 p.

U.S. Geological Survey, 1955–1961, Investigations of Alaskan volcanoes: U.S. Geol. Survey Bull. 1028A–1028T.

U.S. Weather Bureau, 1943, Climatic atlas for Alaska: Army Air Forces Weather Inf. Branch Headquarters, Rept. 44, 229 p.; supp., 72 p.

Vangengeim, E. A., and Sher, A. V., 1970, Siberian equivalents of the Tiraspol faunal complex: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 8, p. 197–207.

Velikovsky, Immanuel, 1955, Earth in upheaval: Garden City, N.Y., Doubleday, 301 p.

Vernon, Petier, and Hughes, O. L., 1966, Surficial geology, Dawson, Larsen Creek and Nash Creek map-areas, Yukon Territory: Canada, Geol. Survey, Bull. 136, 25 p.

Viereck, L. A., 1968, Botanical dating of recent glacial activity in western North America, *in* Osburn, W. H., and Wright, H. E., Jr., Arctic and alpine environments: Bloomington, Indiana Univ. Press, p. 189–204.

Von Huene, R. V., Shor, G. G., and Reimnitz, Erk, 1967, Geological interpretations of seismic profiles in Prince William Sound, Alaska: Geol. Soc. America Bull., v. 78, p. 259–268.

Wahrhaftig, Clyde, 1949, The frost-moved rubble of Jumbo Dome and their significance in the Pleistocene chronology of Alaska: Jour. Geology, v. 57, p. 216–231.

——— 1951, Geology and coal deposits of the western part of the Nenana coal field, Alaska, *in* Barnes, F. F., and others, Coal investigations in south-central Alaska, 1944–1946: U.S. Geol. Survey Bull 963–E, p. 169–186.

——— 1958, Quaternary geology of the Nenana River valley and adjacent parts of the Alaska Range, *in* Wahrhaftig, Clyde, and Black, R. F., Quaternary and engineering geology in the central part of the Alaska Range: U.S. Geol. Survey Prof. Paper 293, p. 1–68.

——— 1965, Physiographic divisions of Alaska: U.S. Geol. Survey Prof. Paper 482, 52 p.

Wahrhaftig, Clyde, and Black, R. F., 1958, Engineering geology along part of the Alaska Railroad, *in* Wahrhaftig, Clyde, and Black, R. F., Quaternary and engineering geology in the central part of the Alaska Range: U.S. Geol. Survey Prof. Paper 293, p. 69–119.

Wahrhaftig, Clyde, and Cox, Allan, 1959, Rock glaciers in the Alaska Range: Geol. Soc. America Bull., v. 70, p. 383–436.

Walker, H. J., 1967, Riverbank dunes in the Colville delta, Alaska: Louisiana State Univ. Coastal Studies Bull. 1 (Tech. Rept. 36), p. 7–14.

Walker, H. J., and Arnborg, L., 1966, Permafrost and ice-wedge effect on riverbank erosion, *in* Permafrost Internat. Conf., Lafayette, Ind., 1963, Proc.: Natl. Acad. Sci.–Natl. Research Council Pub. 1287, p. 164–171.

Walker, H. J., and McCloy, J. M., 1969, Morphologic change in two Arctic deltas: Arctic Inst. North America Research Paper 49, 91 p.

Walker, H. J., and Morgan, H. M., 1961, Morphology of a Colville River delta lake, Alaska—a preliminary report: Alaskan Sci. Conf., 12th, 1961 College, Proc., p. 133.

——— 1964, Unusual weather and river bank erosion in the delta of the Colville River, Alaska: Arctic, v. 17, p. 41–47.

Wallace, R. E., 1948, Cave-in lakes in the Nabesna, Chisana, and Tanana River Valleys, eastern Alaska: Jour. Geology, v. 56, p. 171–181.

Waller, R. M., 1957, Ground water and permafrost at Bethel, Alaska: Alaska Dept. Health, Sec. Sanitation and Eng., Water Hydrol. Data Rept. 2, 11 p.

Ward, W. H., and Sewell, B. A., 1950, Protection of the ground from thermal effects of industrial plant: Geotechnique, v. 2, no. 1, p. 64–81.

Washburn, A. L., 1956, Classification of patterned ground and review of suggested origins: Geol. Soc. America Bull., v. 67, p. 823–866.

Wasowicz, J., 1929, Studies on the snow-line in Canada and Alaska: Acad. Polonaise Sci. Lett. Bull., ser. A, p. 390–399.

Watson, C. E., 1959, Climate of Alaska, *in* Climate of the States: U.S. Weather Bureau, Climatography of the United States no. 60–49, 24 p.

Weber, F. R., and Péwé, T. L., 1961, Engineering geology problems in the Yukon-Koyukuk lowland, Alaska, *in* Short papers in the geologic and hydrologic sciences 1961: U.S. Geol. Survey Prof. Paper 424–D, p. D371–D373.

——— 1970, Surficial and engineering geology of the central part of the Yukon-Koyukuk lowland, Alaska: U.S. Geol. Survey Misc. Geol. Inv. Map I–590, scale 1:125,000.

Weidick, A., 1968, Observations on some Holocene glacier fluctuations in West Greenland: Meddel. om Grønland, band 165, no. 6, p. 1–202.

Wentworth, C. K., and Ray, L. L., 1936, Studies of certain Alaskan glaciers in 1931: Geol. Soc. America Bull., v. 47, p. 879–934.

Wernecke, Livingstone, 1932, Glaciation, depth of frost, and ice veins of Keno Hill and vicinity, Yukon Territory: Eng. and Mining Jour., v. 133, p. 38–43.

Whitmore, F. C., Jr., and Foster, H. L., 1967, *Panthera atrox* (Mammalia: Felidae) from central Alaska: Jour. Paleontology, v. 41, no. 1, p. 247–251.

Wiggins, I. L., 1962, A flora of the Alaskan Arctic Slope: Arctic Inst. North America Spec. Pub. no. 4, 425 p.

Wilcox, R. E., 1959, Some effects of recent volcanic ash falls, with special reference to Alaska: U.S. Geol. Survey Bull. 1028–N, p. 409–476.

——— 1965, Volcanic-ash chronology, *in* Wright, H. E., Jr., and Frey, D. G., eds., The Quaternary of the United States: Princeton, Princeton Univ. Press, p. 807–816.

Wilkerson, A. S., 1932, Some frozen deposits in the gold fields of interior Alaska: Am. Mus. Novitates 525, 22 p.

Williams, J. R., 1955, Yukon Flats, *in* Hopkins, D. M., Karlstrom, T. N. V., and others, Permafrost and ground water in Alaska: U.S. Geol. Survey Prof. Paper 264–F, p. 124–126.

——— 1960, Cenozoic sediments beneath the central Yukon Flats, Alaska, *in* Short papers in the geological sciences 1960: U.S. Geol. Survey Prof. Paper 400–B, p. B329.

——— 1962, Geologic reconnaissance of the Yukon Flats district, Alaska: U.S. Geol. Survey Bull. 1111–H, p. H289–H331.

——— 1970, Ground water in the permafrost regions of Alaska: U.S. Geol. Survey Prof. Paper 696, 83 p.

Williams, J. R., and Ferrians, O. J., Jr., 1961, Late Wisconsin and Recent history of the Matanuska Glacier, Alaska: Arctic, v. 14, p. 82–90.

Williams, J. R., Péwé, T. L., and Paige, R. A., 1959, Geology of the Fairbanks (D–1) quadrangle, Alaska: U.S. Geol. Survey Geol. Quad. Map GQ–124, scale 1:63,360.

Williams, J. R., and Van Everdingen, R. O., 1973, Ground water investigations in permafrost regions of North America—a review, *in* Permafrost, The North American contribution to the Second International Conference, Yakutsk, Siberia, U.S.S.R.: Washington, D.C., Natl. Acad. Sci., p. 435–446.

Winterkorn, H. F., 1943, The condition of water in porous systems: Soil Sci., v. 56, p. 109–115.

Wolfe, J. A., and Leopold, E. B., 1967, Neogene and early Quaternary vegetation of northwestern North America and northwestern Asia, *in* Hopkins, D. M., ed., The Bering Land Bridge: Stanford, Stanford Univ. Press, p. 193–206.

Wood, F. J., ed., 1966, The Prince Williams Sound, Alaska, earthquake of 1964 and aftershocks, v. 1: U.S. Coast and Geod. Survey Pub. 10–3, 263 p.

Wright, H. E., Jr., 1961, Late Pleistocene climate of Europe; a review: Geol. Soc. America Bull., v. 72, p. 933–984.

Yeend, Warren, 1971, Glaciation of the Ray Mountains, central Alaska, *in* Geological Survey research 1971: U.S. Geol. Survey Prof. Paper 750–D, p. D122–D126.

Young, S. B., 1971, The vascular flora of St. Lawrence Island with special reference to floristic zonation in the arctic regions: Harvard Univ., Gray Herbarium, Contr. 201, p. 11–115.

Zeuner, F. E., 1945, The Pleistocene Period: London, Quaritch, 322 p.

INDEX

[Italic page numbers indicate major references]

☆GPO 790-115-1977